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...NOMICAL PROBLEM OF NEUTRAL HELIUM AND HYDROGEN

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AERONOMICAL PROBLEM OF NEUTRAL HELIUM AND HYDROGEN**

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AUTHORS' ABSTRACT

The vertical distribution of helium-4, helium-3, hydrogen-2 and hydrogen-1 in the thermosphere has been investigated, taking account of the various production rates in the mixing region, and different escape rates above the thermopause.

The average escape during the 11-year period from 1951 to 1961 is about 3.5 He^3 atoms $cm^{-2}sec^{-1}$ and 6 x 10^4 He^4 atoms $cm^{-2}sec^{-1}$, corresponding to the atmospheric production of He^3 and to one percent of the generation rate of He^4 in the earth's crust and mantle. The effect of the solar cycle on the escape rates is such that their ratios reach 10^5 for He^4 and 10^4 for He^3 between 1954 and 1958.

The effusion velocities lead, at 500 km, to lifetimes of the order of several years for He^4 (11 years to 1 year for temperatures between 1700° K and 2000° K), of less than 100 days for He^3 for temperatures higher than 1700 K, and of a few hours for H^1 (2 hours to 10 hours for temperatures between 2000 K and 1250 K). These various lifetimes lead to a normal diffusion distribution of He^4 and He^3 in the thermosphere and above the thermopause in almost all circumstances, while the vertical distribution

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of atomic hydrogen and its concentration above the thermopause is subject to variations depending on the temporal variation of the temperature.

The concentration of helium-4 above the thermopause is such that the mean molecular mass at 750 km is less than 16, i.e. less than the mass of atomic oxygen. The helium belt is subject to variations depending on the diffusion level and on the escape flow of atomic hydrogen. Its effective lower boundary reaches lowest altitudes when the thermopause temperature decreases, and its upper bondary reaches highest altitudes when the temperature increases.

The diffusion level of He^4 may vary according to various atmospheric conditions in the 100-120 km region since the descent of the diffusion level from 115 km to 110 km, and from 110 km to 105 km, requires about 15 days and 30 days, respectively. In any case, diffusion conditions related to the level of the turbopause shows that the concentration $n(He)^4$ at 120 km should be $(1 \pm 0.5 \times 10^7$ atoms cm^{-3} leading to exactly a factor of 10 less at 500 km for thermopause temperatures between 2000° K and 1000° K.

The relative concentration of atomic hydrogen in the thermosphere depends on the possivle maximum velocity of diffusion between 100 and 200 km, which is a few cm sec^{-1}. Its vertical distribution approximates a mixing distribution in the lower thermosphere and depends in the upper thermospeere on a scale height limited by ab excape flow almost identical to the maximum flow supported by the diffusion in the mixing region.

If a concentration $n(H) = 10^7$ cm^{-3} is adopted at 100 km, corresponding to a total content of 10^{12} atoms cm^{-2} between 120 km and 200 km, the maximum flow corresponding to the escape rate at 100 km is 2.5×10^7 atoms cm^{-2} sec^{-1}. Under such circumstances, the concentration of atomic hydrogen decreases rapidly above 100 km. It reaches only $10^4 cm^{-3}$ at 500 km for a thermopause temperature of

the order of 1200 K and decreases by a factor of 10 ehen the temperature reaches 2000 K. Since atomic hydrogen reaches a diffusion equilibrium near 500 km, the concentration decreases slowly with height and still is 2.5×10^3 cm^{-3} at 3000 km if the thermopause temperature is 2000 K. Any change in the boundary condition at 100 km, i.e. n (H) = 10^7 cm^{-3}, will modify all parameters dealing with atomic hydrogen by the same factor.

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I. INTRODUCTION

Attention has been recently drawn to the importance of helium and hydrogen in the upper atmosphere. First of all, and because of the importance of its liberation, the problem of helium, re-introduced because of the importance of its escape (cf. Spitzer [37], Mayne [26], Nicolet [28], Bates and McDowell [1], Damon and Kulp [8, 9], Turekian [38], Givarry [14], Byunter [4], has developed on account of its essential influence in the upper part of the heterosphere, as was demonstrated by Nicolet [31, 32]. Among the essential questions emerging are those concerned with the appearance in the atmosphere and the liberation therefrom of the two isotopes He^4 and He^3. Besides, the question of vertical distribution in the thermosphere of these two constituents must be raised. It is indeed dependent upon the diffusion linked with the effect of the thermal diffusion coefficient due to temperature gradient and to the presence in the mixing region of a source. Finally, the physical conditions which are essentially variable at the thermosphere level, must be applied, taking into account the results obtained by Nicolet [33] and Jacchia [19].

It should be immediately added that the problem of hydrogen must be considered in another form, for the various observations made by Purcell & Tousey [35], Chubb, Friedman, Kreplin and Mange [7, 21], show that new parameters must be introduced for a more complete analysis. The basis of the study of vertical distributions in the thermosphere under the influence of a source is given by Mange [23, 25]. Similar studies appear in the paper Harteck and Jensen [16], Urey [39] and Byunter [5]. A complete application to the thermosphere by Bates and Patterson [2] points to possible differences in concentrations at the altitude level of the beginning of diffusion, when the vertical transport and diffusion effects are introduced simultaneously.

Consequently, the aeronomical problem of neutral helium and hydrogen is re-examined taking into account the results obtained on density variations of the heterosphere, associated to altitude and temperature variations of the thermopause.

2. VERTICAL DISTRIBUTION OF TEMPERATURE IN THE HETEROSPHERE

In order to determine with a sufficient precision the physical conditions to which atomic helium and hydrogen are subject, it is appropriate to refer to model of heterosphere where both, the temperature distribution in the thermosphere and the temperature of the thermopause are well defined. By adopting the various conditions specified in detail by Nicolet [33] , one obtains various vertical distributions of temperature compiled in Table 1. A detailed comparison of the density values derived theoretically with the detailed information deduced from satellites has been carried out by Jacchia [19]. We may conclude from that analysis that the noon temperatures may exceed 2000° K at time of solar activity extreme maximum, and they easily reach values approaching 1000° K

at time of solar activity minimum. Moreover, the diurnal temperature variation of the thermpause corresponds to 300° K differences for the lowest values and easily reaches 500° K for the highest values. These variations stem from both, the importance of heating by the solar ultraviolet and the rapidity of heat conductivity. As to the magnetic storm effect, its interpretation is also given by a general temperature increase (Nicolet [30]), for Jacchia has shown [18], that the density increases at all latitudes, and Dessler [10], Fransis & Karplus [13] have shown that the hydromagnetic heating is general.

In order to make more precise the exact conditions at which the study of the beginning of diffusion is made, Table II presents the parameters utilized between 100 and 120 km.

3. LIBERATION OF GASES.

When dealing with the problem of gases' liberation, it is practical to refer to the formulae utilized by Nicolet [28]. Thus, the number of atoms F_c, liberated per unit of time and of surface at the critical level r_c is given by :

$$F_c = n_c \, (g_c/2\,\pi)^{1/2} \, r_c \, H_c^{-1/2} \, (1 + H_c/r_c) \, e^{-r_c/H_c} \qquad (1)$$

In this formula, the critical level, situated at the distance r_c from the center of the Earth, corresponds to an altitude where it is admitted that the number of collisions is no longer sufficient to allow every atom having attained the liberation speed to escape from the atmosphere. The index c indicates that all the symbols utilized refer to the critical level : g_c is the gravitation acceleration, H_c is the height of a scale defined by $H_c \equiv k\,T_c/m\,g_c$ where k is the Boltzmann constant, if T_c is the absolute temperature in °K amd m — the atom mass considered expressed in g.

It must be understood that in order to obtain this formula, we considered that the atoms' velocity corresponded to a Maxwellian distribution.

If one admits that the vertical distribution of the constituent above the critical level may be represented by a diffusion equilibrium distribution in an isothermal atmosphere, the following formula may be adopted for the concentration n_a of an atom of mass m at a level $r = a$ situated at a distance z_c below the critical level r_c:

$$n_c = n_a \exp\left(-\frac{a}{a + z_c} \cdot \frac{z_c}{H_a}\right) \qquad (2)$$

Taking advantage of (2), formula (1) may be written:

$$F_c = n_a (g_a/2\pi)^{1/2} \frac{a}{1 + z_c/a} \cdot \frac{1 + H_a\,(1 + z_c/a)/a}{H_a^{1/2}\; e^{a/H_a}} \qquad (3)$$

Adopting the 500 km altitude, i.e. $a = 6870$ km, we may write (3) in a practical form, F_c cm^{-2} sec^{-1},

$$F_c = 8{,}16 \times 10^9\, n_{500}\, \frac{1 + H_{500}/6{,}87 \times 10^8}{H_{500}^{1/2}\; e^{6{,}87 \times 10^8/H_{500}}} \qquad (4)$$

where the ratio $z_c / 6.87 \times 10^8$ is neglected ahead of the unity. The adoption of the 500 km altitude is very convenient, for it allows to avoid the exact definition of the critical level, by merely making a slight error, and it permits also the adoption of a diffusion equilibrium in any case.

One may write that

$$v_E = F_c / n_{500} \qquad (5)$$

in other words, that by dividing in formula (4) the flux F_c by by the concentration n_{500}, one obtains the expression for the rate of effusion at the critical level. Calculation shows (see Table III) that the effusion rates for atoms of mass M = 1 (atomic hydrogen) to M = 4 (helium-4) are quite different. A temperature of 1600°K leads respectively to effusion rates of 3.6×10^{-2} cm sec^{-1}, 2.3×10^{-1} cm sec^{-1}, 1.5×10^{2} cm sec^{-1} and 9.7×10^{3} cm sec^{-1} for He^4, He^3, H^2 and H^1. Figures 1, 2 and 3 show how the effusion rates vary as a function of temperature.

If we admit first of all, that the concentrations above 500 km are fixed by a diffusion equilibrium, it is possible to determine the lifetime of a given atom subject to a release effect. Defining the lifetime τ_E as the time required to reduce the concentration to 50% of the equilibrium value, and considering that

$$n_a H_a Y = \int_a^\infty n \, dz,$$

i.e. the total quantity above a certain level r = a, we obtain by means of formula (5) :

$$\tau_E = \frac{0{,}693 \int_z^\infty n \, dz}{n_z v_E} \qquad (6)$$

or, at the 500 km altitude,

$$\tau_E \text{ (500 km)} = \frac{0{,}693 \, H_{500} \, Y}{v_E} \qquad (7)$$

Factor Y, computed by utilizing the Herring and Kyle method [17] and represented in the Fig. 4, shows the possible order of magnitude of the error when utilizing $nH = n\,dz$ at the exosphere level. For helium the error cannot exceed 20 percent. Thus, Fig. 5 represents the minimum lifetime of He^4 at 500 km, computed with the aid of formula (6) for temperatures from 1600 to 2400° K. One may see that for such temperature interval the variation is superior to the factor of 100. Thus, 50 days at 2350° K would be equivalent to 5000 days at 1650° K.

If we then admit, that helium-4 diffusion begins at 105 km, we may seek to ascertain what the lifetime of all helium content in the atmosphere would be. By adopting the value of 1.13×10^{20} atoms of $He^4\,cm^{-2}$ for the atmosphere content, we find lifetimes indicated in Fig. 6. A temperature variation from 1150 to 2250° K makes the He^4 lifetime pass from 2.5×10^9 years to 2.5×10^5 years. Such variation shows indeed that no conclusions may be drawn concerning a mean atmosphere temperature. Furthermore, one must avoid to concede absolute values to these lifetimes. Comparison of the lifetimes represented in Fig. 6 whith those derived by Damon and Kulp [9] points to a difference by a factor greater than 10. In fact these lifetimes of total helium content depend on the level of the beginning of diffusion. If in particular the diffusion level were situated at 110 or 115 km, instead of 105 km introduced for the calculation, lifetimes of Fig. 6 ought to be multiplied by approximately 2 or 4. That is why it is preferable to examine the question by introducing a representation on the conditions of release by the study of conditions inherent to specific temperature at 500 km and then to find out what the evolution as a function of temperature is. Fig. 7 presents an indication on the time ratio required to obtain a specific release of the same number of helium-4 atoms.

The variation is such that is necessary to know with great precision the thermopause temperatures and the corresponding time intervals. In other words, because of thermopause temperature variations, it is impossible to determine a priori what should the permanent state be, wherever the atom He^4 production globally compensates the losses by escape.

In the case of He^3, the minimum lifetime at 500 km varies from 10 to 10 000 days for temperatures diminishing from 2150°K to 1250°K (see Fig. 8). Since it is necessary to maintain temperatures beyond 1700°K in order that the concentration of He^3 at 500 km be reduced by one half in less than 100 days, one may realize that the equilibrium conditions of He^3 diffusion correspond indeed to normal conditions of that constituent. Only under special circumstances, corresponding to highest temperatures related to a very high solar activity during a period sufficently long, could a reduction of He^3 concentration in the thermosphere take place. At that moment (T 2100°K), the diffusion could no longer support the flux of He^3 atoms escaping from the atmosphere, and the concentration of He^3 at 500 km would be reduced by one half in some ten days, (see Figures 8 and 10).

Thus, considering that temperatures above 2000°K appear only under exceptional circumstances, we may admit that the vertical distribution of He^4 and He^3, submitted to diffusion, is normally a distribution of diffusion equilibrium.

One may thus admit that the concentration at 500 km is fixed in the thermosphere by the diffusion equilibrium as of a certain level called the level of the beginning of diffusion. It is thus possible to utilize formula (4). The application made for He^4 and He^3 with equilibrium of diffusion beginning at the 105 km level, where the respective concentrations are

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$$n_{105}(He^4) = 3.00 \times 10^7 \text{ cm}^{-3} \quad (8)$$

$$n_{105}(He^3) = 3.75 \times 10^1 \text{ cm}^{-3} \quad (9)$$

lead to the following values at 500 km depending on temperature

$$n_{500}(He^4) = (2 \pm 0.3) \times 10^6 \text{ cm}^{-3} \quad (10)$$

$$n_{500}(He^3) = (3.4 \pm 0.2) \text{ cm}^{-3} \quad (11)$$

Such concentrations lead to escape fluxes represented in Fig. 9 and 10, we may see immediately that even for highest temperatures, the ratio of helium-4 to helium-3 escape never reaches the corresponding ratio for concentrations observed in the troposphere. That is why the issue of general economy of the different atoms in the whole of the atmosphere must be raised.

4. GENERATION OF HE^4 AND HE^3 AND THEIR PRODUCTION IN THE ATMOSPHERE.

Whatever are the contemplated methods for the determination of helium-4, one obtains always values greater than 10^6 atoms $cm^{-2} sec^{-1}$ for its generation in the terrestrial globe. Nicolet determined a generation of $1.75 \times 10^6 cm^{-2} sec^{-1}$ particles α for an amount corresponding to a quantity of 9.65 g cm^{-2} of uranium and of 30.55 g cm^{-2} of thorium [28]. A similar value of 2.5×10^6 $cm^{-2} sec^{-1}$ may be derived from Mayne's calculations [26] by utilizing the production of 3 particles α per gram of granit utilized by Birch [3]. The maximum generation of helium-4 given by Damon and Kulp [9] and which 7×10^6 particles α $cm^{-2} sec^{-1}$, corresponds to the importance attributed to the generation in the mantle, and which is of the order of 5×10^6 $cm^{-2} sec^{-1}$. The viewpoints adopted by Mayne [26] and Damon and Kulp [9] differ. The first adopts a generation of helium-3 in the atmosphere that

could be compatible with the generation of helium-4, while the second authors consider a retention of helium-4 in the crust and in the mantle that could be compatible with a specific production of helium-3. By adopting an atmospheric production of He^3 of the order of 2 $cm^{-2}sec^{-1}$, Damon and Kulp [9] are compelled to conclude that, taking into account a ratio $n(He^3)/n(He^4) = 1.2 \times 10^{-6}$, the atmospheric production of He^4 is only $4.5 \times 10^4 cm^{-2}sec^{-1}$. However, the temperature of at least 1880° K they determined is certainly too high, as this is indicated by the values of Fig. 9. It would be indeed necessary to diminish by a factor of 20 the concentration of He^4 adopted at 500 km for the escape calculation plotted in Figure 9. In any case, less than one percent of He^4 produced in the interior of the Earth would have been released in the atmosphere under the form indicated by Damon and Kulp [9].

Another method introduced by Turekian [38] offers only a very limited interest. By adapting the atmospheric production of He^4 to that of argon generated by the radioactive potassium K^{40}, this author derives a He^4 production of about $5.4 \times 10^5 cm^{-2}sec^{-1}$ which he identifies with an equivalent escape erroneously requiring a thermopause temperature of the order of 2200° K. The equivalent production of He^3 ought to be of the order of 10 atoms $cm^{-2}sec^{-1}$. One may see that this last value corresponds to that of Mayne [26] which he introduced attempting to justify it by an accretion of extra-terrestrial dust. It would be appropriate to recall here the Fireman determinations [12] giving He^3 productions as standing between 0.7 and 1.5 $cm^{-2}sec^{-1}$ and justifying values of the order of 2 $cm^{-2}sec^{-1}$ with a possible error of ±50%.

However, if it is clear that the escape of He^3 and He^4 depends on the thermopause temperature, one nevertheless cannot link their generation and atmospheric production only on the basis of that unique parameter for which a mean value would be adopted.

Such stand leads to erroneous results either in the atmospheric production of He^3 or in the determination of the retention inside the globe of He^4. Moreover, it is clear that the temperatures one may adopt for the thermopause are erroneous when one fails to account for the structure of the thermosphere within which helium and hydrogen diffuse. Finally, the very pronouced effect of the very great temperature variation as shown in Figure 7, must be taken into account. One may see that the transition from 1500 to 2000°K allows the release of the same number of He^4 atoms in a time 100 times shorter. Thus, the real escape can be only determined by attributing a specific weight for certain temperatures. This may only be made by analyzing the variation of thermopause temperature in the course of one solar activity cycle.

5. DIFFUSION IN THE THERMOSPHERE.

It is evident that the concentration of various elements at the critical level depends on the conditions at which the diffusion is effected in the thermosphere. First of all, the boundary conditions make it necessary to select a level of the beginning of diffusion. It is expedient to refer first of all to the observation data based upon the results obtained with the aid of mass-spectrographs. The Johnson's critical analysis [20] proves that credit may be given to determinations of inert gases' ratio. Such is the case of the ratio $n(A)/n(N_2)$ which leads to an observation at the beginning of diffusion. Referring to Meadows and Townsend [27] and to Pokhunkov [34], one may admit that diffusion certainly appears in the region (110 ± 5) km. Taking up the analysis of the diffusion for a minor constituent, one may say that the rate of diffusion w_1 for a constituent of mass m_1 and of

concentration n_1 is expressed by

$$w_1 = - D_1\left[\frac{1}{n_1} \frac{\partial n_1}{\partial z} + \frac{1}{H} \left\{ \frac{m_1}{m} + \left(\beta - \frac{2H}{a+z} \right) (1 + \alpha_T) \right\} \right] \quad (12)$$

where H is the height of atmospheric scale with a gradient β represented by an equivalent average constituent of mass m and concentration $\underline{n}$. $\underline{a}$ is the terrestrial radius at the level $z = 0$, and α_T is the thermal diffusion coefficient (Chapman and Cowling [6]). The factor D_1 multiplying the bracket in (12) is the diffusion coefficient which is written in the following form :

$$D_1 = \frac{3\sqrt{\pi/2}}{8\pi\sigma_{12}^2} \left(1 + \frac{m_2}{m_1}\right)^{1/2} (gH_2)^{1/2} \cdot \frac{1}{n_2} \quad (13)$$

where σ_{12} is the collision diameter.

Considering that the continuity equation

$$\frac{\partial n_1}{\partial t} + \frac{\partial}{\partial z} (n_1 w_1) = 0 \quad (14)$$

where w_1 is represented by (12), is applicable to the region from 100 to 120 km (see Table II), one may determine the diffusion times ihe sense developped by Mange [24]. The application may be made by the following formula, defining the diffusion time τ_D:

$$\tau_D = A\, H_2^2 / D_1 \quad (15)$$

where A is a constant to be determined. This formula supplies immediately the following relative times :

Altitude (km)	120	115	110	105	100
D	1	1.58	2.62	4.62	8.89

If we now consider the existence of a source or of a leakage in the atmosphere, we write that at a height z above the level of the sphere of radius a

$$n_1 w_1 = S = S_a \left(\frac{a}{a+z}\right)^2 \qquad (16)$$

so as to take into account the condition of continuity. Introducing into (16) the expression for w_1 where H_1 — the scale height of the minor constituent — is substituted for H_2 — the atmospheric scale height, we thus write the following relation for a stationary state :

$$\frac{\partial n_1}{\partial z} + \frac{n_1}{H_1}\left[1 + \left(\beta_1 - \frac{2H_1}{a+z}\right)\left(1 + \alpha_T\right)\right] + \frac{S_a}{D_1}\left(\frac{a}{a+z}\right)^2 = 0 \qquad (17)$$

If the value of the 3rd term of (17) is sufficiently low, the constituent of mass m_1 is in diffusion equilibrium, since the 2 first terms correspond to the experssion (12) where the diffusion rate is zero. In other words, there exist values of flux S_a which are easily withstood by the diffusion to allow the equilibrium, and there exists a maximum value of S beyond which the diffusion equilibrium may no longer be applied in the vertical distribution.

The diffusion flux S is directly derived from the expression (12) for the diffusion rate. Considering that the atmospheric constituent likely to diffuse is a perfect mixing, one may postulate that in (12)

$$\frac{1}{n_1}\frac{\partial n_1}{\partial z} \equiv \frac{1}{n}\frac{\partial n}{\partial z} \qquad (18)$$

since the distribution of the constituent of concentration n_1 follows the distribution of the atmospheric constituent of concentration n.

Introducing (18) and (13) into (12), we obtain for $S = n_1 w_1$

$$S = \frac{3\sqrt{\pi/2}}{8\pi\,\sigma_{12}^2}\left(1+\frac{m}{m_1}\right)^{1/2}\left(\frac{m}{k}\right)^{1/2}\frac{g n_1}{T^{1/2}\,n}\left[1-\frac{m_1}{m}-\alpha_T\left(\beta-\frac{2H}{a+z}\right)\right] \qquad (19)$$

Taking into account the general conditions stipulated in Table II, we obtain for the flux S supplied by the diffusion at 100 km, the following values :

$$S_{100}\,(A^{40}) = -3{,}6 \times 10^{10}\ cm^{-2}sec^{-1} \qquad (20a)$$

$$S_{100}\,(He^{4}) = 8{,}8 \times 10^{7}\ cm^{-2}sec^{-1} \qquad (20b)$$

$$S_{100}\,(He^{3}) = 1{,}3 \times 10^{2}\ cm^{-2}sec^{-1} \qquad (20c)$$

$$S_{100}\,(H^{1}) = 2{,}5 \times 10^{7}\ cm^{-2}sec^{-1} \qquad (21)$$

The application of formula (19) is made taking into account that a mean factor of thermal diffusion α_T may be adopted :

$$\alpha_T = -0.4 \qquad (22)$$

The conditions adopted for the helium are presented in Table IV and correspond to the general conditions of Table I. As to hydrogen, the absolute value of diffusion flux cannot be determined directly if the concentration under the diffusion region is not fixed. We first admit that even in case of dissociation, the vertical distribution at 100 km is not dependent upon the photochemical equilibrium.

The time required by the mixing is shorter than that required to reach the photochemical equilibrium. Since the presence of hydrogen is assured by the molecule H_2 (relative abundance — 5×10^{-7} per volume in the homosphere), it must thus be admitted that in the case of dissociation of molecules in a mixing region, the relative abundance of the hydrogen atom may reach 10^{-6} per volume, and so the concentration of atomic hydrogen is of the order of

$$n_{100\,km}^{(H)} \simeq 10^7\ cm^{-3} \qquad (23)$$

The concentration of hydrogen indicated by (23) is adopted for the calculation of flux given by (21). By adopting a possible diffusion effect with a flux S of the order of $2.5 \times 10^7 cm^{-2} sec^{-1}$ we find after integration of the equation (17) the values of the concentration n(H). Figure 11 represents the vertical distribution of the atomic hydrogen from 100 to 200 km under the conditions fixed by the atmospheric parameters of Tables I and II. We may see that the vertical distribution of atomic hydrogen from 100 to 200 km remains practically the same in the two cases respectively corresponding to the perfect mixing and to the diffusion. The conditions are such that there is little difference in the concentrations of hydrogen obtained at 120 km no matter what limit is assigned at the turbopause, i.e. at the end of the mixing zone.

6. DIFFUSION AND RELEASE (ESCAPE) OF HELIUM-4.

It is expedient to recall that in diffusion equilibrium the concentrations of He^4 at 500 km are obtained by taking into account solely the beginning of diffusion. Considering the temperature distributions presented in Table I and taking as diffusion start the levels between 105 and 120 km, we obtain the various values compiled in Table V. It may be seen at once (Table VI and Fig. 9) that He^4 escape always corresponds to a flux lower than

that given by formula (20b), even for highest temperatures and concentrations. In fact, since there is question of limitation imposed in the homosphere, one may see that He^4 concentration in the heterosphere is neither limited by a release at the critical level, nor by a maximum diffusion rate. The condition fixing a limit is the level of the beginning of diffusion, i.e. the effective altitude of the turbopause. As to the atmospheric production of He^4, it is linked with the retention of particles inside the terrestrial globe.

In conclusion, one may admit that the beginning of diffusion, placed at 105 km, represents the optimum condition leading to the greatest abundance of He^4 in the heterosphere. In any case, a reduction to 1/10 of He^4 concentrations obtained with a beginning of diffusion at 105 km would require a beginning of diffusion beyond 120 km.

7. DIFFUSION AND RELEASE OF HE^3

By applying to He^3 conditions analogus to those of He^4, i.e. diffusion equilibrium for the vertical distribution beginning at a specific level, one determines the conditions described by the data of Table VII for a beginning of diffusion at 105 km. We may see immediately (Table VII and Figure 10) that for the highest temperatures ($T > 2100^\circ K$), we would obtain liberation or escape flux values superior to those that the diffusion can endure (formula 20 c). Consequently, the concentration represented in Table VII for $T = 2133^\circ K$ is de facto a fictive concentration which is reduced by one half in 10 days. But, as was already stressed above, temperatures above 2100° K would be exceptional temperatures, appearing only during rather brief periods. Thus, under normal conditions, the diffusion equilibrium represents the normal

distribution of He^3 in the thermosphere where the vertical distribution of that element is subject to the diffusion phenomenon.

8. PROBLEM OF THE BEGINNING OF HELIUM DIFFUSION.

Taking up the Mange criterion defined by the formula (15) which gives the diffusion time, and determining in that formula the constant A for argon and helium, it appears immediately that the diffusion is about three times faster for argon than for helium. The time necessary to bring down the starting level of the diffusion equilibrium is function of altitude. It is given in Table VIII for altitudes from 120 to 105 km and a lowering corresponding to an altitude interval of 5 km. Besides, Figure 12 describes the time required to bring the beginning of diffusion equilibrium to altitudes between 115 and 100 km when the starting level is placed 5 km above. One may, thus see to what extent the necessary times to lower the starting level of the diffusion equilibrium increase when that level is farther and farther down. Since the concentration of helium at 120 km, when the diffusion equilibrium may be assured rapidly, will depend on the vertical distribution of He^4 below that altitude, it is expedient to determine what the absolu- concentrations of He^4 may be, according to the various possibilities that may be vizualized. First of all, below the region where the diffusion equilibrium exists, i.e. below 120 km, the distribution of helium will depend upon the miximg conditions of the atmosphere. If the turbopause, that is to say the limit where the mixing is no longer continually maintained, is fixed, one may determine the value of the concentration at 120 km and adopt for that calculation a beginning of equilibrium of the diffusion. We remind in Table IX of n(He) concentrations which we would obtain at 120 km by admitting various levels for the diffusion equilibrium. These values indicate that a factor of 10 separates He^4 concentrations at 120 km

depending on whether one adopts a beginning of diffusion equilibrium at 120 km or at 100 km. In fact, the value $n(He^4)$ at 120 km depends on the turbopause level. Concentrations of He^4 are represented in Fig. 13 by the various conditions that may appear. It may be seen how the concentration varies as a function of the turbopause level for diffusion equilibrium beginnings respectively fixed at o05 km, 110 km and 115 km. Thus, a value of He^4 concentration may be established with a certain approximation, and it is given by

$$n_{120\,km}(He^4) = (1.25 \pm 0.75) \times 10^7 \text{ cm}^{-3} \qquad (24)$$

The maximum value may thus be identified for a diffusion equilibrium beginning equivalent to 105 km, corresponding to a turbopause altitude below 100 km. The minimum value requires a beginning of diffusion equilibrium above 115 km with a permanent turbopause above 105 km. Thus it appears that a beginning of diffusion equilibrium established at 110 km, may be viewed as acceptable conditions for a computation. The adopting of 105 km instead of 110 km does not entail an increase in helium concentrations in the heterosphere greater than by a factor of 2.

In conclusion, it may be admitted that starting with the value $n(He^4) \simeq 10^7$ cm^{-3} at 120 km, it is possible to compute the vertical distribution of that element in all the heterosphere.

9. VERTICAL DISTRIBUTION OF ATOMIC HYDROGEN.

The vertical distribution of atomic hydrogen concentration has been determined by formula (17); it is established first of all in the lower thermosphere by the flux the diffusion can support. This distribution does not differ much from that existing in the case of perfect mixing, as this is shown by the $n(H)$ representation in Fig. 11. When the effect of diffusion is evident at one or another level below 120 km, the concentration at 100 km is always established to a 10 percent approximation, i.e. with the absolute values consi-

dered by us, $(1.0 \pm 0.1) \times 10^7$ hydrogen atoms at 100 km thus correspond to a sources of 2.5×10^7 $cm^{-2}sec^{-1}$ in the homosphere.

The computation with the aid of formula (17) shows that the altitude of 500 km must be reached to allow the utilization of diffusion equilibrium in the determination of the vertical distribution of n(H). It results therefrom that the scale height associated to the representation of atomic hydrogen vertical distribution between the altitudes of 100 and 500 km varies by the value of the atmospheric scale height — $H \simeq 10$ km, at the hydrogen scale height $H > 1000$ km. Fig.11 has already shown that the vertical distribution of the hydrogen atom varies between 100 and 200 km. The vertical distribution of hydrogen in the lower thermosphere is not very different from that of the principal constituents of the atmosphere. It is expedient to emphasize this, for, as Bates and Patterson have already shown that in their computations [2], the extrapolation toward lower altitudes starting from 500 km cannot be effected the way numerous authors have done it*). Fig.14 represents the true vertical distribution of atomic hydrogen between 100 and 500 km, which is compared with the one that would be derived from diffusion equilibrium. For that purpose, the 500 km level is taken as the starting point for the comparison and the relative concentration $n(H) = 1$ is taken at 100 km. We may see at once to what extent the extrapolation by diffusion equilibrium starting at 500 km may lead to erroneous concentrations in the thermosphere.

Since the diffusion equilibrium is applicable beyond 500 km, it is possible to determine the lifetime of a hydrogen atom prior to its release. The application of formula (7) leads to the values of Table X, where it may be seen, that for temperatures $> 1000^\circ K$, the concentrations may vary in the course of a 24-hour period.

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*) At the IIIrd Cospar Symposium in Washington D. C., May 1962, Kallmann, Krasovskiy, Harris and Priester, and Paetzold have also presented vertical distribution of hydrogen in diffusion equilibrium.

In other words, as a consequence of the diurnal temperature variation, the concentration of hydrogen at 500 km must vary. Fig.15 represents this variation of n(H), dropping to $10^3 cm^{-3}$ for the conditions of strong solar activity, and increasing to $10^5 cm^{-3}$ during quiet Sun periods. Finally, it is clear that such variations should appear between day and night on account of diurnal variation of thermopause temperature.

Since hydrogen concentration beyond the thermopause varies very strongly as a function of temperature, it is expedient to examine what the distribution of the total number of molecules in specific vertical columns can be. First of all, the total number of hydrogen atoms between 100 and 120 km being practically fixed by the conditions presented in Table II, we obtain a nearly constant value :

$$\int_{100\,km}^{120\,km} n\,dz = (5.65 \pm 0.10) \times 10^{12}\ cm^{-2} \qquad (25)$$

Then, the variation of the content of the column from 120 to 200 km does not change much and is adapted to Purcell and Tousey observations [35] and to Bates and Patterson analysis [2].

$$\int_{120\,km}^{200\,km} n\,dz = (1.05 \pm 0.20) \times 10^{12} cm^{-2} \qquad (26)$$

It is clear that repeated observations in this atmospheric region should permit an excellent calibration of the content in atomic hydrogen. A marked variation appears between 200 and 500 km, which is naturally manifest between 500 km and 3000 km. Table XI indicates what the variation is as a function of thermopause temperature. It thus corresponds to the calibration defined by one or the other relation (23) or (25) or (26). But it should be stressed, that these values correspond to a stationary state, which will be

reached with difficulty, for the effect of diurnal temperature variation will result in a constant modification of vertical distribution. As to the total number of hydrogen atoms beyond 3000 km, it represents from ten to twenty percent of the total determined beyond 3000 km. Fig.16 represents the different total numbers of hydrogen atoms in the different layers considered above. Thus it appears that the hydrogen content beyond 200 km varies very strongly when the temperature varies between 1000° K and 1800° K. Such differences ought to be detectable by adequate observations.

In order to show the difference in the behavior of helium and hydrogen, Fig.17 presents the vertical distribution of these two elements between 500 km and 3000 km, First of all, the concentration of He^4 at 500 km does not practically depend on temperature, while that of H^1 is distinctly different. Then, the differences between the concentrations of He^4 due to temperature become more and more important as the altitude increases, while those of H^1 do not vary. Besides, helium concentration increases with rising temperatures, while that of hydrogen diminishes. Finally, helium prevails in the whole of that atmosphere region so long as the temperature is above 1000° K. However, it is expedient to mention that the absolute values of concentrations of hydrogen, as well as of helium at 500 km depend upon the absolute values adopted in the low lower thermosphere. The curve shift in Fig.17 may be effected by bringing farther or nearer the concentrations of He and H. These curves are indeed based upon the values $n(H) = 10^7 cm^{-3}$ at 100 km, and $n(He) = 2 \times 10^7 cm^{-3}$ at 120 km.

In short, the vertical distribution of atomic hydrogen is very strongly dependent upon temperature, and an analysis of the problem of Lyman-α radiation, such as that of Donahue [11], points to the importance of the knowledge of boundary conditions

in the thermosphere before determining exactly the distribution of neutral hydrogen at greater distances.

10. PROBLEM OF DEUTERIUM.

If helium is immediately in the state of diffusion equilibrium when the effect of diffusion is manifest without perturbation, and if hydrogen-1 cannot reach a diffusion equilibrium except at high altitudes, one may wonder what the behavior of hydrogen-2 is.

Calculation shows (see Fig.18), that the diffusion equilibrium level appears to be lower for the deuterium than for hydrogen-1 and for lowest temperatures in particular. The result of this is that the extreme concentrations n(D) at 500 km vary by about a factor of 45, while those of hydrogen-1 — n(H) vary only by a factor of 37. Nevertheless, the decrease of n(D) with altitude is much less rapid than that of n(H). We may indeed see that

$$n(D)_{500\,km} > 10^{-2}\ n(D)_{100\,km} \quad \text{for} \quad T < 1800^{\circ}K,$$

while $$n(H)_{500\,km} < 10^{-2}\ n(H)_{100\,km} \quad \text{for} \quad T > 900^{\circ}K.$$

In other words, there is a pronounced variation of concentrations n(H) and n(D) in the thermosphere as a function of temperature, while that of helium — $n(He^4)$ does not even vary by a factor of 1.5 at 500 km. But the essential difference between the normal and the heavy hydrogen is evidenced by a variation of the ratio n(H)/n(D) as a function of altitude. Fig.19 shows how this ratio can vary. The various curves related to different thermopause temperatures show that it is difficult to draw precise conclusions, except of course for a decrease between 100 and 200 km of the ratio n(H)/n(D), so much the more rapid as the temperature drops. Above 200 km the variation is such that it is difficult to conclude without referring to the lower region. In any case, one may expect for a large temperature range a decrease to 5×10^{-2} of the ratio n(H)/n(D) from 100 to 500 km.

11. RELEASE OF GASES IN THE COURSE OF A SOLAR ACTIVITY CYCLE

In order to have a complete outlook on helium escape, it is expedient to follow the variation in the course of a complete cycle of solar activity, taking also into account the diurnal variation. The period 1951 — 1961 has been chosen since we are in possession of complete data on radioelectric solar flux, and of partial data since 1957 on satellite period variation associated with the variation of temperature. As a matter of example, the flux of He^3 and He^4 atoms, released per cm^2 sec, is presented in Table XII for 1954 (solar activity minimum) and for 1958 (solar activity maximum) during daytime. One may see immediately the enormous difference between a quiet sun year and a year of intense solar activity. A factor of 10^5 appears between the escape flows of He^4. Considering the 11-year period from 1951 to 1961, we obtain a mean value $F_c(He^3)$ for the escape flow of helium-3.

$$F_c(He^3)_{1951-1961} = 3.5 \text{ atoms } cm^{-2}sec^{-1} \qquad (27)$$

i.e. a value of the same order of magnitude than that derived from the production by cosmic radiation (see paragraph 4). Since this value is related to fixing at the beginning of diffusion, at 105 km, we are certain to find in any case $F_c(He^3) = (2 \pm 1)\, cm^{-2}sec^{-1}$.

Under the same conditions, we obtain for He^4

$$F_c(He^4)_{1951-1961} = 6 \times 10^4 \text{ atoms } cm^{-2}sec^{-1} \qquad (28)$$

i.e. only a value of the order of one percent of the total generation of helium-4. This weak escape of He^4 would correspond to a retention of some 99 percent of helium-4 in the mantle and the earth's crust and could be justified if the escape of helium-4 from the earth's crust is very slow.

Since the various analyses [15], [22], [36], [26], [8], [38], tending to determine the percentage of helium-4 produced in the earth's crust that passes into the atmosphere, cannot but lead to indeterminateness, the solution of the problem of He^4 release cannot be made more precise. If it is clear, for example that the photoionization of He^4 may easily lead to 2×10^6 ions $He^+ cm^{-2} \cdot sec^{-1}$ [32], we must also figure the escaping processus either by direct path, corresponding to the passage through the magnetic pole, or by indirect path, permitting to the ion ahead of the neutral atom to have a sufficient velocity. In the case of the ion escaping directly, it is difficult to imagine the sufficiently rapid passage of ions from low to polar latitudes by way of horizontal diffusion. In the case of a transformation of the ion into a neutral atom with a sufficient kinetic energy, we must obviously admit a total efficiency. Nevertheless, this processus must not influence the He^3 ions, for otherwise there would be too little He^3 in the atmosphere.

In fact, the result obtained for He^3 in the course of a solar activity cycle, indicating the agreement between the production and escape of that element, leads to the conclusion that a very important retention of He^4 generated in the earth's crust by radioactivity, might exist. However, an escape cause influencing He^4 and not He^3 is not excluded, and it results thereform that the quantitative data concerning He^4 escape are difficult to obtain.

12. Atmospherical Concentrations from 500 km to 3000 km.

Taking up the determinations of thermospherical concentrations of nitrogen, oxygen, helium and hydrogen, based on temperature distribution in Table I, on physical parameters of Table II, the concentration of helium-4 of Table V with diffusion at 105 km and a flux of 2.5×10^7 hydrogen atoms $cm^{-2} sec^{-1}$ at 100 km, we obtain the concentrations presented in Table XIII.

The molecular mass variation from 500 km, where $M = 17 \pm 2$, to 3000 km, where $M = 2.5 \pm 1.5$, shows how the constitution of the heterosphere successively passes from a molecular nitrogen — atomic oxygen composition to that of helium-hydrogen, passing by an atmosphere exclusively composed of helium. At high temperatures, helium still remains the essential element, even at 3000 km, and hydrogen does not play any important part, except when temperature approaches 1000° K.

It is however expedient to emphasize before closing several addtional points. First of all, the distribution of atomic oxygen may always be considered in hydrostatic equilibrium to greatest distances, for there is no escape for that element. Neutral helium may be influenced at a sufficient distance from the center of the Earth, since there is release. In case of hydrogen we may consider that its distribution to 3000 km does not differ a great deal from the hydrostatic distribution. Therefore, data of Table XIII correspond to a normal distribution of all the different constituents based upon the conditions established at the basis of the isothermal layer.

The case of hydrogen distribution corresponds to a stationary case, where no perturbation of any sort is introduced. However, any variation, rapidly revealed at 500 km (a few hours). reaches easily the region of the lower thermosphere, say the 100 — 120 km altitude range, in one or two days.

Hence, the problem of vertical distribution of hydrogen cannot be resolved completely without considering simultaneously its variation as a function of time (diurnal variation), and the conditions at boundaries (its production below 100 km).

It will thus be necessary to proceed with numerous observations of the constitution of the heterosphere, where helium and hydrogen intervene as principal elements. Variations will appear, that will have to be put in parallel with those of thermopause temperature. They will thus be related to variations of solar flux and to atmosphere cooling by conduction.

***** E N D *****

References, Tables and Figures follow..

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TABLE I.- Vertical distribution of temperatures ($^{\circ}$K) for various temperatures of the thermopause

Altitude (km)	1	2	3	4	5	6	7	8	9
130	509	509	509	509	505	516	515	511	505
140	694	694	696	695	688	677	664	649	634
150	877	873	863	852	829	803	777	751	726
160	1059	1022	1002	982	941	902	864	827	793
170	1212	1148	1117	1088	1031	979	930	884	842
180	1340	1252	1212	1174	1104	1039	981	927	878
190	1448	1340	1292	1246	1162	1087	1020	960	905
200	1540	1414	1358	1305	1210	1126	1051	985	925
210	1618	1477	1413	1355	1248	1156	1075	1004	940
220	1685	1530	1460	1396	1280	1180	1094	1018	951
240	1791	1613	1532	1458	1327	1215	1120	1037	965
260	1870	1674	1585	1503	1358	1237	1136	1048	973
280	1930	1719	1622	1534	1379	1252	1145	1054	977
300	1975	1752	1650	1556	1393	1261	1150	1057	978
320	2010	1777	1670	1592	1402	1267	1153	1059	
340	2036	1796	1685	1583	1408	1271	1155		
360	2057	1809	1695	1590	1411	1272			
380	2073	1819	1702	1594	1412				
400	2086	1826	1707	1597					
420	2096	1831	1710	1598					
440	2104	1835	1711						
460	2110	1837							
480	2115								
500	2119								

TABLE II

TABLEAU II.- : Paramètres physiques de 100 à 120 km. (Physical parameters from 100 to 120 km)

z (km)	T (°K)	H (cm)	$n(N_2)$ (cm^{-3})	$n(O_2)$ (cm^{-3})	n(O) (cm^{-3})	n(M) (cm^{-3})	ρ (gm cm^{-3})
100	200	$6,37 \times 10^{5}$	$1,08 \times 10^{13}$	$2,19 \times 10^{12}$	$1,41 \times 10^{12}$	$1,44 \times 10^{13}$	$6,55 \times 10^{-10}$
105	231	$7,37 \times 10^{5}$	$4,50 \times 10^{12}$	$9,15 \times 10^{11}$	$5,87 \times 10^{11}$	$6,00 \times 10^{12}$	$2,73 \times 10^{-10}$
110	262	$8,37 \times 10^{5}$	$2,10 \times 10^{12}$	$4,28 \times 10^{11}$	$2,74 \times 10^{11}$	$2,81 \times 10^{12}$	$1,28 \times 10^{-10}$
115	293	$9,37 \times 10^{5}$	$1,07 \times 10^{12}$	$2,17 \times 10^{11}$	$1,39 \times 10^{11}$	$1,43 \times 10^{12}$	$6,49 \times 10^{-11}$
120	324	$10,37 \times 10^{5}$	$5,82 \times 10^{11}$	$1,18 \times 10^{11}$	$7,60 \times 10^{10}$	$7,77 \times 10^{11}$	$3,54 \times 10^{-11}$

TABLE III Effusion velocity V_E (cm sec^{-1}) as a function of the temperature of the thermopause.

T (°K)	He^4	He^3	H^2	H^1
2133	$2,45 \times 10^{0}$	$5,46 \times 10^{1}$	$1,24 \times 10^{3}$	$2,66 \times 10^{4}$
1837	$3,24 \times 10^{-1}$	$1,20 \times 10^{1}$	$4,57 \times 10^{2}$	$1,64 \times 10^{4}$
1711	$1,09 \times 10^{-1}$	$5,34 \times 10^{0}$	$2,68 \times 10^{2}$	$1,26 \times 10^{4}$
1598	$3,56 \times 10^{-2}$	$2,31 \times 10^{0}$	$1,54 \times 10^{2}$	$9,66 \times 10^{3}$
1412	$3,78 \times 10^{-3}$	$4,31 \times 10^{-1}$	$5,10 \times 10^{1}$	$5,63 \times 10^{3}$
1273	$4,51 \times 10^{-4}$	$8,78 \times 10^{-2}$	$1,78 \times 10^{1}$	$3,38 \times 10^{3}$
1155	$5,12 \times 10^{-5}$	$1,72 \times 10^{-2}$	$6,07 \times 10^{0}$	$2,00 \times 10^{3}$
1059	$5,98 \times 10^{-6}$	$3,44 \times 10^{-3}$	$2,09 \times 10^{0}$	$1,19 \times 10^{3}$
978	$7,02 \times 10^{-7}$	$6,89 \times 10^{-4}$	$7,24 \times 10^{-1}$	$7,07 \times 10^{2}$

TABLEAU IV.- Concentrations (cm^{-3}) de He^4 et de He^3 dans l'atmosphère sans effet de diffusion.

TABLE IV. Concentration (in cm^{-3}) of He^4 and He^3 in the atmosphere without diffusion.

Altitude	100 km	105 km	110 km	115 km	120 km
He^4	$7,2 \times 10^7$	$3,0 \times 10^7$	$1,4 \times 10^7$	$7,2 \times 10^6$	$3,9 \times 10^6$
He^3	$9,00 \times 10^1$	$3,75 \times 10^1$	$1,75 \times 10^1$	9,00	4,88

TABLE V. Concentrations (cm^{-3}) of He^4 at 500 km in case of diffusion equilibrium beginning at 105, 110, 115 and 120 km

Temperature	Diffusion beginning at			
(°K)	105 km	110 km	115 km	120 km
2133	$2,3 \times 10^6$	$1,3 \times 10^6$	$7,3 \times 10^5$	$4,5 \times 10^5$
1837	2,2	1,2	7,2	4,5
1711	2,2	1,2	7,1	4,4
1598	2,2	1,2	7,1	4,4
1412	2,1	1,2	6,8	4,2
1273	2,0	1,1	6,5	4,0
1155	1,9	1,0	6,1	3,8
1059	1,8	$9,8 \times 10^5$	5,7	3,5
978	1,7	9,2	5,3	3,3

TABLEAU VI.- Libération de He^4 (F_c cm^{-2} sec^{-1}) avec début de diffusion débutant à 105, 110, 115 et 120 km.

TABLE VI (Liberation of He^4 ($F_c cm^{-2} sec^{-1}$) with diffusion beginning at the indicated altitudes)

Température (°K)	Diffusion commençant à 105 km	110 km	115 km	120 km
2133	$5,43 \times 10^6$	$3,0 \times 10^6$	$1,8 \times 10^6$	$1,1 \times 10^6$
1837	$7,41 \times 10^5$	$4,1 \times 10^5$	$2,4 \times 10^5$	$1,5 \times 10^5$
1711	$2,51 \times 10^5$	$1,4 \times 10^5$	$8,1 \times 10^4$	$5,0 \times 10^4$
1598	$7,83 \times 10^4$	$4,4 \times 10^4$	$2,5 \times 10^4$	$1,6 \times 10^4$
1412	$8,19 \times 10^3$	$4,6 \times 10^3$	$2,6 \times 10^3$	$1,6 \times 10^3$
1273	$9,58 \times 10^2$	$5,3 \times 10^2$	$3,1 \times 10^2$	$1,9 \times 10^2$
1155	$9,71 \times 10^1$	$5,4 \times 10^1$	$3,1 \times 10^1$	$1,9 \times 10^1$
1059	$1,13 \times 10^1$	$6,3 \times 10^0$	$3,6 \times 10^0$	$2,3 \times 10^0$
978	$1,25 \times 10^0$	$6,9 \times 10^{-1}$	$4,0 \times 10^{-1}$	$2,5 \times 10^{-1}$

TABLE VII Concentrations (cm^{-3}) of He^3 at 500 km and escape (cm^{-2} sec^{-1}) in case of diffusion equilibrium beginning at 105 km

Température (°K)	$n(He^3)$ (cm^{-3})	F_c ($cm\ sec^{-1}$)
2133	3,7	205
1837	3,6	47
1711	3,6	21
1598	3,6	9
1412	3,6	2
1273	3,5	$3,5 \times 10^{-1}$
1155	3,5	$6,6 \times 10^{-2}$
1059	3,3	$1,3 \times 10^{-2}$
978	3,2	$1,5 \times 10^{-3}$

TABLEAU VIII.- Temps nécessaire pour abaisser le début du niveau d'équilibre de diffusion de 5 km.

TABLE VIII (Time necessary to lower the beginning of the diffusion equilibrium level by 5 km)

Altitude (km)	A^{40}	He^4	Altitude (km)
A partir de 120	3,5 jours	10,0 jours	atteignant 115
A partir de 115	5,8	16,6	atteignant 110
A partir de 110	10,3	29,3	atteignant 105
A partir de 105	20	56	atteignant 100

TABLE IX. Concentrations of He^4 at 120 km for various altitudes of beginning of diffusion

Altitude (km)	$n(He^4)$ (cm^{-3})	Altitude (km)	$n(He^4)$ (cm^{-3})
120	$3,9 \times 10^6$	110	$1,1 \times 10^7$
117,5	$5,0 \times 10^6$	107,5	$1,4 \times 10^7$
115	$6,3 \times 10^6$	105	$1,9 \times 10^7$
112,5	$8,0 \times 10^6$	102,5	$2,7 \times 10^7$
		100	$3,9 \times 10^7$

TABLEAU X.- Concentrations et durées de vie de l'hydrogène atomique à 500 km; (concentration $n(H)_{100\ km} = 10^7\ cm^{-3}$)

TABLE X (Concentration and lifetime of atomic hydrogen at 500 km)

T(thermopause) (°K)	n(H) (cm^{-3})	τ (heures) (hours)
2133	$8,3 \times 10^2$	2
1837	$1,3 \times 10^3$	3
1711	$1,7 \times 10^3$	4
1598	$2,3 \times 10^3$	5
1412	$3,9 \times 10^3$	7
1273	$6,5 \times 10^3$	10
1155	$1,1 \times 10^4$	16
1059	$1,9 \times 10^4$	24
978	$3,1 \times 10^4$	37

TABLEAU XI.- Nombre total d'atomes d'hydrogène (cm^{-2}) dans les colonnes verticales de 200 à 500 km et de 500 à 3000 km. Calibration $n(H)_{100\ km} = 10^7\ cm^{-3}$.

TABLE XI. (Total number of H atoms in the vertical columns from 200 and 500 km and from 500 to 3000 km)

T (thermopause) (°K)	200-500 (km)	500-3000 (km)
2133	$1,3 \times 10^{11}$	$1,3 \times 10^{11}$
1837	$1,5 \times 10^{11}$	$2,0 \times 10^{11}$
1711	$1,6 \times 10^{11}$	$2,5 \times 10^{11}$
1598	$1,7 \times 10^{11}$	$3,2 \times 10^{11}$
1412	$2,2 \times 10^{11}$	$5,1 \times 10^{11}$
1273	$3,0 \times 10^{11}$	$7,9 \times 10^{11}$
1155	$4,5 \times 10^{11}$	$1,3 \times 10^{12}$
1059	$7,1 \times 10^{11}$	$2,0 \times 10^{12}$
978	$1,2 \times 10^{12}$	$3,2 \times 10^{12}$

TABLEAU XII.- Libération de He^3 et He^4 en 1954 et 1958. (He^3 and He^4 escape in 1954 and 1958)

Valeur moyenne mensuelle du nombre d'atomes s'échappant cm^{-2} sec^{-1} pendant le jour. (Mean monthly value of the number of escaping atoms , in cm^{-2} sec^{-1}) in daytime.

		1954		1958	
Mois (Month)		He^3	He^4	He^3	He^4
Jan.	Janvier	$5,6 \times 10^{-3}$	4,5	$3,8 \times 10^{1}$	$7,8 \times 10^{5}$
Feb.	Février	5,6	4,5	1,1	1,2
Mar.	Mars	7,6	6,6	4,2	8,4
Apr.	Avril	5,6	4,5	3,7	7,0
May	Mai	5,6	4,5	1,6	2,2
June	Juin	5,6	4,5	1,3	1,6
July	Juillet	5,6	4,5	2,0	3,1
Aug.	Août	5,6	4,5	2,1	2,9
Sept.	Septembre	5,6	4,5	2,6	3,9
Oct.	Octobre	7,6	6,6	2,0	3,2
Nov.	Novembre	7,7	6,9	1,2	1,6
Dec.	Décembre	9,2	8,8	2,1	2,9

TABLE XIII. Concentration (cm^{-3}), Mass (O = 16) and total density at 3000 km for various temperature of the thermopause.

TABLEAU XIII.- Concentrations (cm^{-3}), masse (O = 16) et densité totale ($gm\ cm^{-3}$) de 500 km à 3000 km pour diverses températures de la thermopause.

Température (°K)	2133	1837	1711	1598	1412	1273	1155	1059	978
500 km									
n(O)	$1,58\times10^{8}$	$1,25\times10^{8}$	$1,10\times10^{8}$	$9,45\times10^{7}$	$6,91\times10^{7}$	$4,96\times10^{7}$	$3,49\times10^{7}$	$2,42\times10^{7}$	$1,63\times10^{7}$
$n(N_2)$	$4,85\times10^{7}$	$2,90\times10^{7}$	$2,17\times10^{7}$	$1,59\times10^{7}$	$8,39\times10^{6}$	$4,35\times10^{6}$	$2,18\times10^{6}$	$1,08\times10^{6}$	$5,17\times10^{5}$
$n(He^4)$	$2,26\times10^{6}$	$2,25\times10^{6}$	$2,20\times10^{6}$	$2,19\times10^{6}$	$2,11\times10^{6}$	$2,01\times10^{6}$	$1,90\times10^{6}$	$1,77\times10^{6}$	$1,66\times10^{6}$
n(H)	$8,27\times10^{2}$	$1,34\times10^{3}$	$1,74\times10^{3}$	$2,28\times10^{3}$	$3,91\times10^{3}$	$6,51\times10^{3}$	$1,10\times10^{4}$	$1,85\times10^{4}$	$3,11\times10^{4}$
Masse	18,9	18,2	17,9	17,6	17,0	16,5	16,1	15,7	15,3
Densité totale	$6,71\times10^{-15}$	$4,8\times10^{-15}$	$4,0\times10^{-15}$	$3,3\times10^{-15}$	$2,3\times10^{-15}$	$1,5\times10^{-15}$	$1,0\times10^{-15}$	$7,1\times10^{-16}$	$4,7\times10^{-16}$
750 km									
n(O)	$2,45\times10^{7}$	$1,49\times10^{7}$	$1,11\times10^{7}$	$8,17\times10^{6}$	$4,32\times10^{6}$	$2,30\times10^{6}$	$1,18\times10^{6}$	$6,00\times10^{5}$	$2,98\times10^{5}$
$n(N_2)$	$1,84\times10^{6}$	$6,98\times10^{5}$	$3,96\times10^{5}$	$2,20\times10^{5}$	$6,58\times10^{4}$	$2,01\times10^{4}$	$5,80\times10^{3}$	$1,68\times10^{3}$	$4,71\times10^{2}$
$n(He^4)$	$1,42\times10^{6}$	$1,32\times10^{6}$	$1,24\times10^{6}$	$1,19\times10^{6}$	$1,06\times10^{6}$	$9,33\times10^{5}$	$8,15\times10^{5}$	$7,04\times10^{5}$	$6,08\times10^{5}$
n(H)	$7,36\times10^{2}$	$1,17\times10^{3}$	$1,51\times10^{3}$	$1,95\times10^{3}$	$3,28\times10^{3}$	$5,37\times10^{3}$	$8,89\times10^{3}$	$1,47\times10^{4}$	$2,42\times10^{4}$
Masse	16,2	15,6	15,2	14,8	13,8	12,6	11,1	9,5	7,8
Densité totale	$7,5\times10^{-16}$	$4,4\times10^{-16}$	$3,2\times10^{-16}$	$2,4\times10^{-16}$	$1,2\times10^{-16}$	$6,8\times10^{-17}$	$3,7\times10^{-17}$	$2,1\times10^{-17}$	$1,2\times10^{-17}$
1000 km									
n(O)	$6,18\times10^{6}$	$2,05\times10^{6}$	$1,32\times10^{6}$	$8,34\times10^{5}$	$3,27\times10^{5}$	$1,31\times10^{5}$	$4,99\times10^{4}$	$1,92\times10^{4}$	$7,15\times10^{3}$
$n(N_2)$	$1,65\times10^{5}$	$2,17\times10^{4}$	$9,45\times10^{3}$	$4,05\times10^{3}$	$7,18\times10^{2}$	$1,34\times10^{2}$	$2,31\times10^{1}$	$4,04\times10^{0}$	--
$n(He^4)$	$9,26\times10^{5}$	$8,03\times10^{5}$	$7,29\times10^{5}$	$6,71\times10^{5}$	$5,53\times10^{5}$	$4,56\times10^{5}$	$3,70\times10^{5}$	$2,97\times10^{5}$	$2,39\times10^{5}$
n(H)	$6,60\times10^{2}$	$1,03\times10^{3}$	$1,32\times10^{3}$	$1,69\times10^{3}$	$2,79\times10^{3}$	$4,48\times10^{3}$	$7,28\times10^{3}$	$1,18\times10^{4}$	$1,91\times10^{4}$
Masse	14,2	12,7	11,8	10,7	8,5	6,6	5,4	4,6	4,1
Densité totale	$1,3\times10^{-16}$	$6,1\times10^{-17}$	$4,0\times10^{-17}$	$2,7\times10^{-17}$	$1,2\times10^{-17}$	$6,5\times10^{-18}$	$3,8\times10^{-18}$	$2,5\times10^{-18}$	$1,8\times10^{-18}$
1250 km									
n(O)	$8,97\times10^{5}$	$3,21\times10^{5}$	$1,80\times10^{5}$	$9,85\times10^{4}$	$2,93\times10^{4}$	$9,00\times10^{3}$	$2,61\times10^{3}$	$7,67\times10^{2}$	$2,20\times10^{2}$
$n(N_2)$	$5,64\times10^{3}$	$8,47\times10^{2}$	$2,90\times10^{2}$	$9,64\times10^{1}$	$1,05\times10^{1}$	$1,24\times10^{0}$	--	--	--
$n(He^4)$	$6,21\times10^{5}$	$5,05\times10^{5}$	$4,43\times10^{5}$	$3,93\times10^{5}$	$3,03\times10^{5}$	$2,33\times10^{5}$	$1,77\times10^{5}$	$1,33\times10^{5}$	$1,00\times10^{5}$
n(H)	$5,97\times10^{2}$	$9,2\times10^{2}$	$1,16\times10^{3}$	$1,48\times10^{3}$	$2,40\times10^{3}$	$3,79\times10^{3}$	$6,05\times10^{3}$	$9,64\times10^{3}$	$1,54\times10^{4}$
Masse	11,1	8,8	7,5	6,4	5,0	4,4	4,1	3,9	3,6
Densité totale	$2,8\times10^{-17}$	$1,2\times10^{-17}$	$7,7\times10^{-18}$	$5,2\times10^{-18}$	$2,8\times10^{-18}$	$1,8\times10^{-18}$	$1,3\times10^{-18}$	$9,2\times10^{-19}$	$7,0\times10^{-19}$
1500 km									
n(O)	$2,01\times10^{5}$	$5,65\times10^{4}$	$2,79\times10^{4}$	$1,33\times10^{4}$	$3,06\times10^{3}$	$7,34\times10^{2}$	$1,65\times10^{2}$	$3,76\times10^{1}$	$8,43\times10^{0}$
$n(N_2)$	$4,12\times10^{2}$	$4,06\times10^{1}$	$1,20\times10^{1}$	$2,91\times10^{0}$	--	--	--	--	--
$n(He^4)$	$4,27\times10^{5}$	$3,27\times10^{5}$	$2,78\times10^{5}$	$2,39\times10^{5}$	$1,72\times10^{5}$	$1,25\times10^{5}$	$8,86\times10^{4}$	$6,25\times10^{4}$	$4,43\times10^{4}$
n(H)	$5,43\times10^{2}$	$8,25\times10^{2}$	$1,03\times10^{3}$	$1,30\times10^{3}$	$2,08\times10^{3}$	$3,23\times10^{3}$	$5,09\times10^{3}$	$7,97\times10^{3}$	$1,25\times10^{4}$
Masse	7,8	5,8	5,1	4,6	4,2	4,0	3,9	3,7	3,3
Densité totale	$8,2\times10^{-18}$	$3,7\times10^{-18}$	$2,6\times10^{-18}$	$1,9\times10^{-18}$	$1,2\times10^{-18}$	$8,5\times10^{-19}$	$6,0\times10^{-19}$	$4,3\times10^{-19}$	$3,2\times10^{-19}$

cont'd

TABLEAU XIII.- (suite) Concentrations (cm^{-3}), masse (O = 16) et densité totale ($gm\ cm^{-3}$) de 500 km à 3000 km pour diverses températures de la thermopause.

TABLE XIII (.. continued)

Température (°K)	2133	1837	1711	1598	1412	1273	1155	1059	978
2000 km									
n(O)	$1,32 \times 10^{4}$	$3,40 \times 10^{3}$	$9,34 \times 10^{2}$	$3,40 \times 10^{2}$	$5,00 \times 10^{1}$	$7,65 \times 10^{0}$	$1,08 \times 10^{0}$	--	--
$n(N_2)$	--	--	--	--	--	--	--	--	--
$n(He^4)$	$2,16 \times 10^{5}$	$1,48 \times 10^{5}$	$1,19 \times 10^{5}$	$9,63 \times 10^{4}$	$6,14 \times 10^{4}$	$3,98 \times 10^{4}$	$2,52 \times 10^{4}$	$1,59 \times 10^{4}$	$1,00 \times 10^{4}$
n(H)	$4,57 \times 10^{2}$	$6,76 \times 10^{2}$	$8,35 \times 10^{2}$	$1,04 \times 10^{3}$	$1,60 \times 10^{3}$	$2,43 \times 10^{3}$	$3,71 \times 10^{3}$	$5,64 \times 10^{3}$	$8,60 \times 10^{3}$
Masse	4,7	4,2	4,1	4,0	3,9	3,8	3,6	3,2	2,6
Densité totale	$1,8 \times 10^{-18}$	$1,1 \times 10^{-18}$	$8,2 \times 10^{-19}$	$6,5 \times 10^{-19}$	$4,1 \times 10^{-19}$	$2,7 \times 10^{-19}$	$1,7 \times 10^{-19}$	$1,1 \times 10^{-19}$	$8,1 \times 10^{-20}$
2500 km									
n(O)	$1,29 \times 10^{3}$	$1,45 \times 10^{2}$	$4,60 \times 10^{1}$	$1,38 \times 10^{1}$	$1,30 \times 10^{0}$	--	--	--	--
$n(N_2)$	--	--	--	--	--	--	--	--	--
$n(He^4)$	$1,18 \times 10^{5}$	$7,35 \times 10^{4}$	$5,59 \times 10^{4}$	$4,29 \times 10^{4}$	$2,46 \times 10^{4}$	$1,45 \times 10^{4}$	$8,24 \times 10^{3}$	$4,69 \times 10^{3}$	$2,69 \times 10^{3}$
n(H)	$3,92 \times 10^{2}$	$5,67 \times 10^{2}$	$6,91 \times 10^{2}$	$8,47 \times 10^{2}$	$1,27 \times 10^{3}$	$1,88 \times 10^{3}$	$2,80 \times 10^{3}$	$4,15 \times 10^{3}$	$6,17 \times 10^{3}$
Masse	4,1	4,0	4,0	4,0	3,8	3,7	3,2	2,6	1,9
Densité totale	$8,2 \times 10^{-19}$	$4,9 \times 10^{-19}$	$3,7 \times 10^{-19}$	$2,9 \times 10^{-19}$	$1,7 \times 10^{-19}$	$9,9 \times 10^{-20}$	$5,9 \times 10^{-20}$	$3,8 \times 10^{-20}$	$2,8 \times 10^{-20}$
3000 km									
n(O)	$1,37 \times 10^{2}$	$1,12 \times 10^{1}$	$3,12 \times 10^{0}$	--	--	--	--	--	--
$n(N_2)$	--	--	--	--	--	--	--	--	--
$n(He^4)$	$6,88 \times 10^{4}$	$3,93 \times 10^{4}$	$2,85 \times 10^{4}$	$2,09 \times 10^{4}$	$1,09 \times 10^{4}$	$5,85 \times 10^{3}$	$3,04 \times 10^{3}$	$1,58 \times 10^{3}$	$8,25 \times 10^{2}$
n(H)	$3,42 \times 10^{2}$	$4,84 \times 10^{2}$	$5,83 \times 10^{2}$	$7,06 \times 10^{2}$	$1,04 \times 10^{3}$	$1,50 \times 10^{3}$	$2,18 \times 10^{3}$	$3,16 \times 10^{3}$	$4,59 \times 10^{3}$
Masse	4,0	4,0	3,9	3,9	3,7	3,4	2,8	2,0	1,5
Densité totale	$4,6 \times 10^{-19}$	$2,6 \times 10^{-19}$	$1,9 \times 10^{-19}$	$1,4 \times 10^{-19}$	$7,4 \times 10^{-20}$	$4,1 \times 10^{-20}$	$2,4 \times 10^{-20}$	$1,6 \times 10^{-20}$	$1,3 \times 10^{-20}$

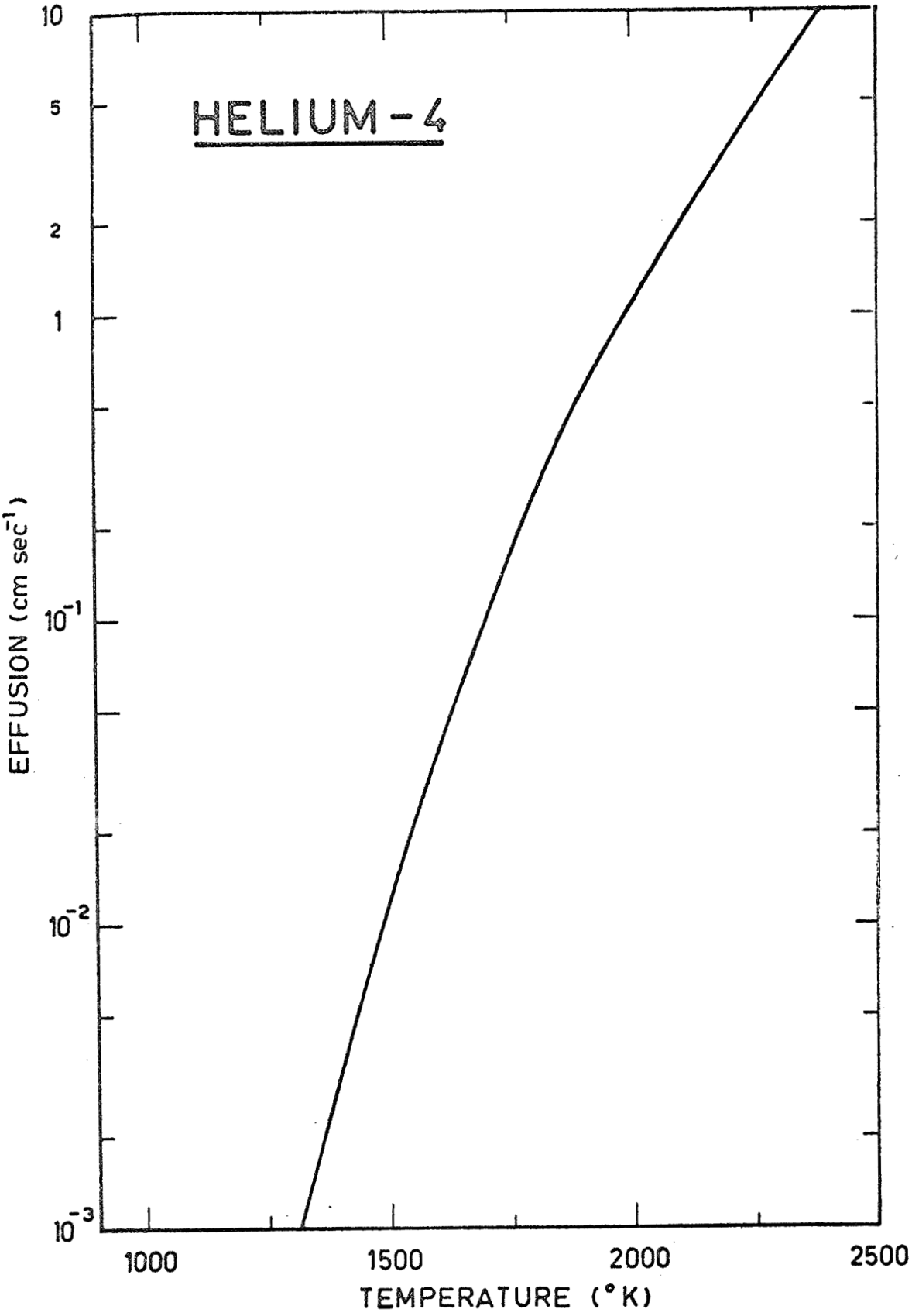

Fig. 1.- Effusion velocity of helium-4 as a function of thermopause temperature, (Rate of Effusion)

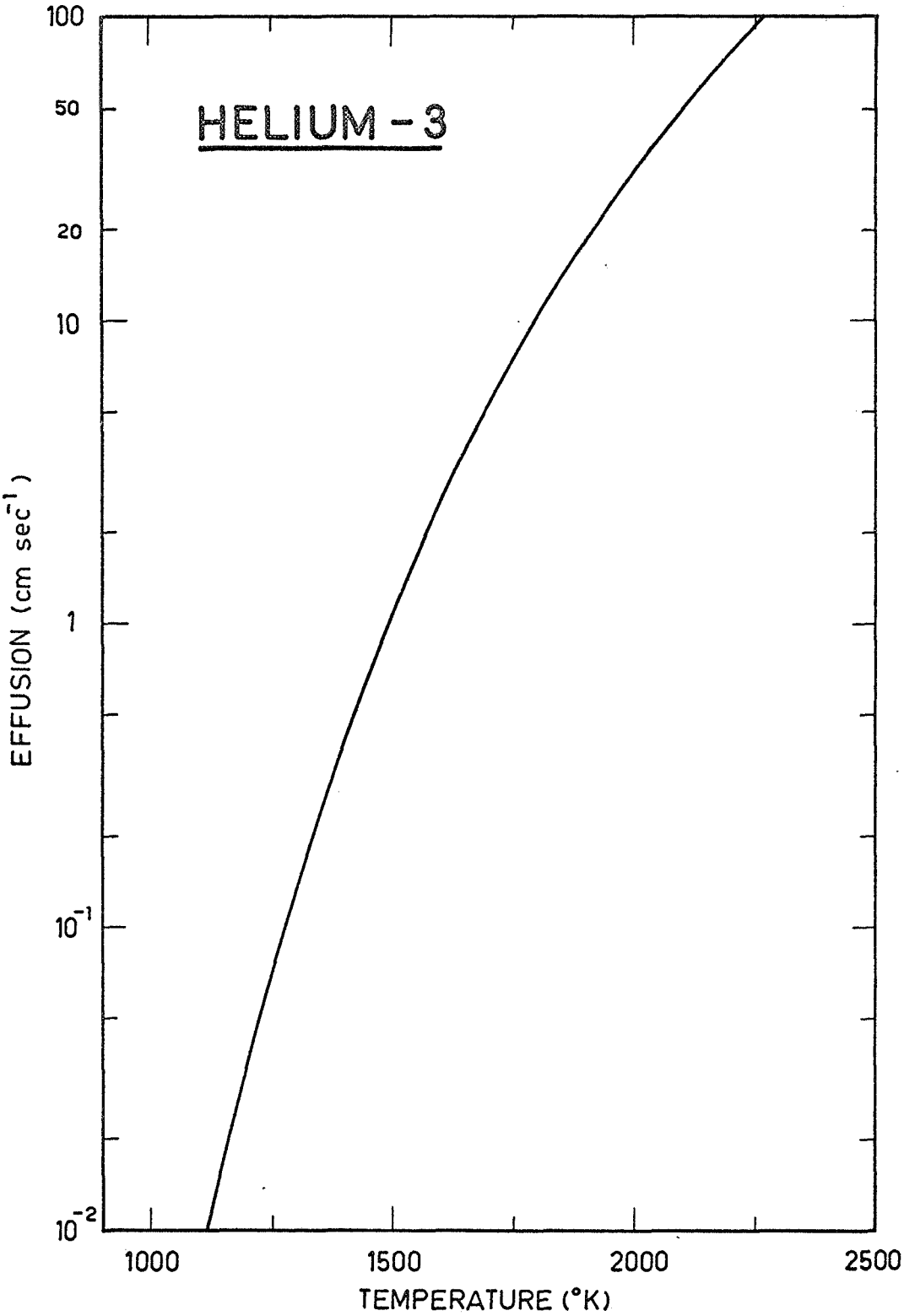

Fig. 2.- Effusion velocity of helium-3 as a function of thermopause temperature. (Rate of Effusion)

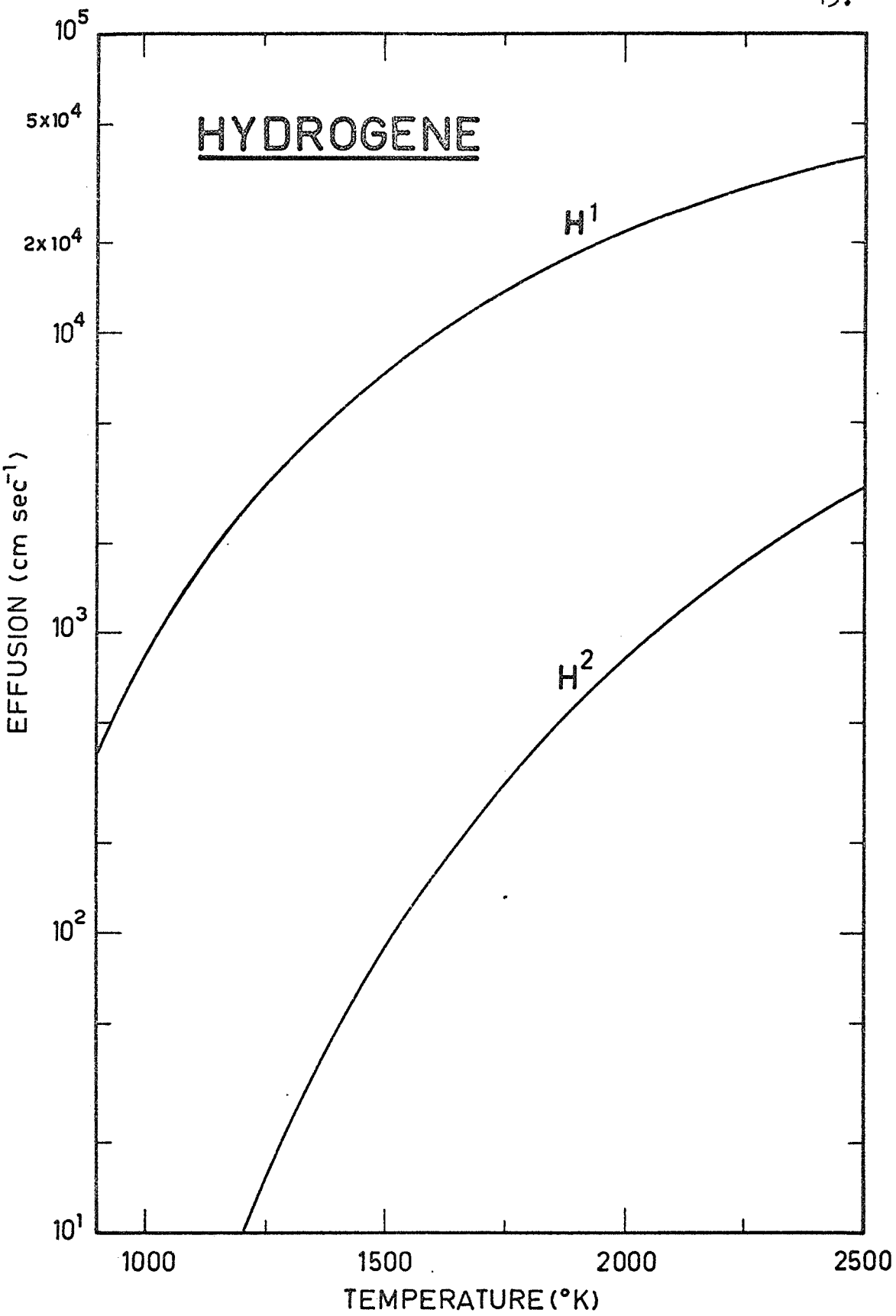

Fig. 3.- V_f Effusion velocity of hydrogen-1 and of hydrogen-2 as a function of the thermopause temperature. (rate of effusion)

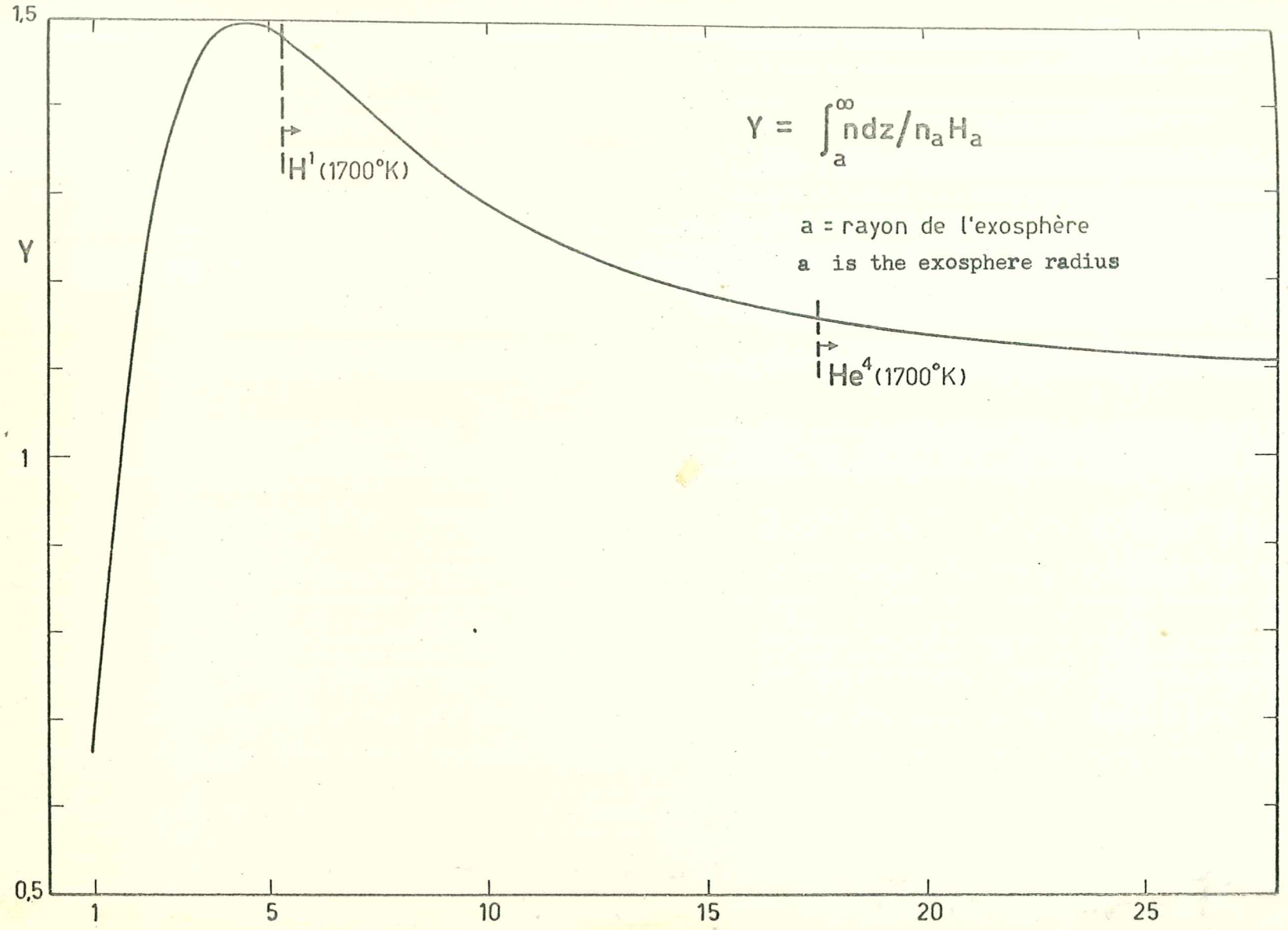

Fig. 4.- Ratio of the total number of atoms starting at the level of exosphere beginning to the number of atoms defined by the product of concentration by the height of the scale nH as a function of the ratio of the critical level radius to the scale height of the considered constituent.

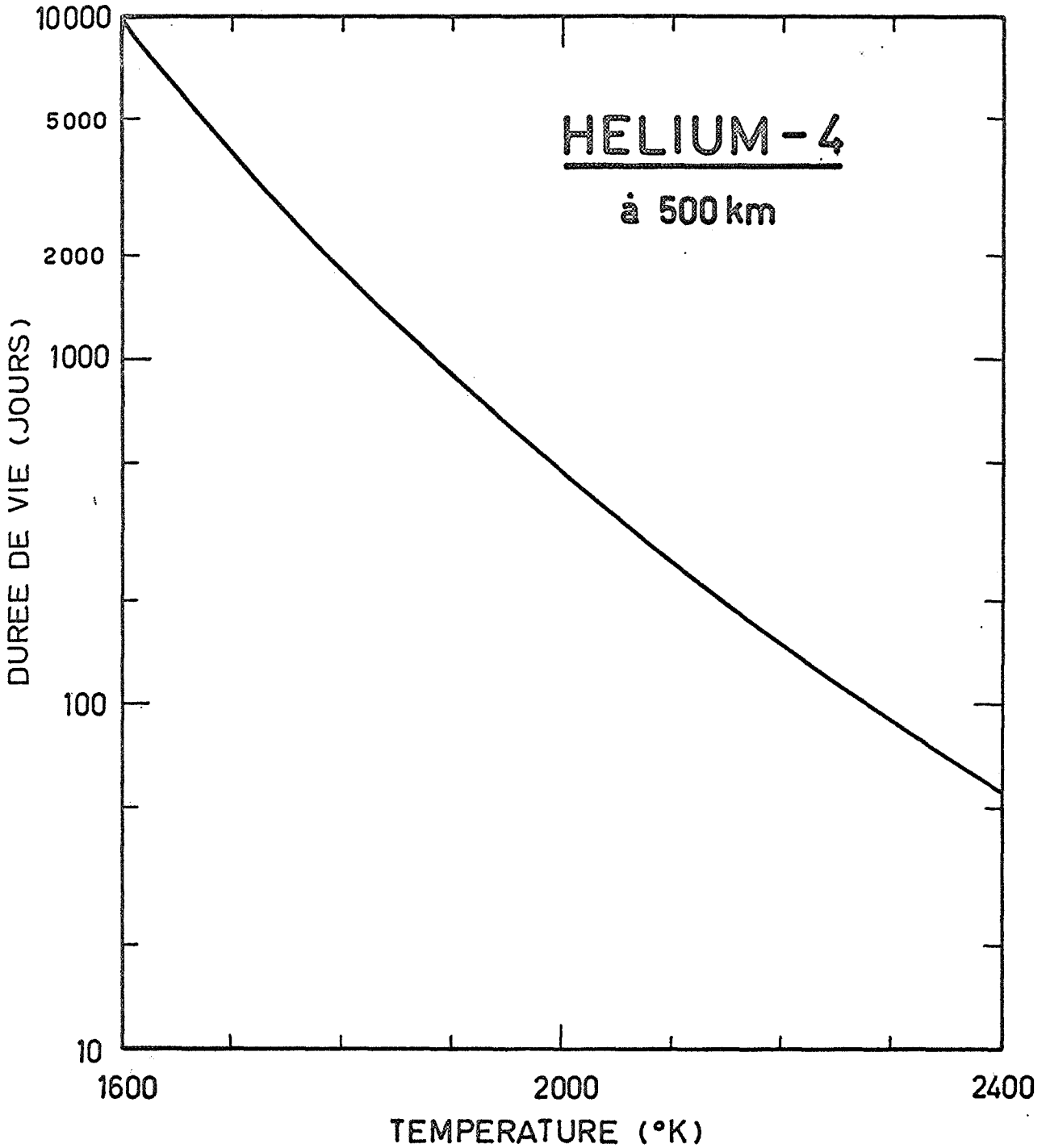

Fig. 5.- Lifetime of a He^4 atom at 500 km as a function of thermopause temperature. Reduction to 50% of the initial value.

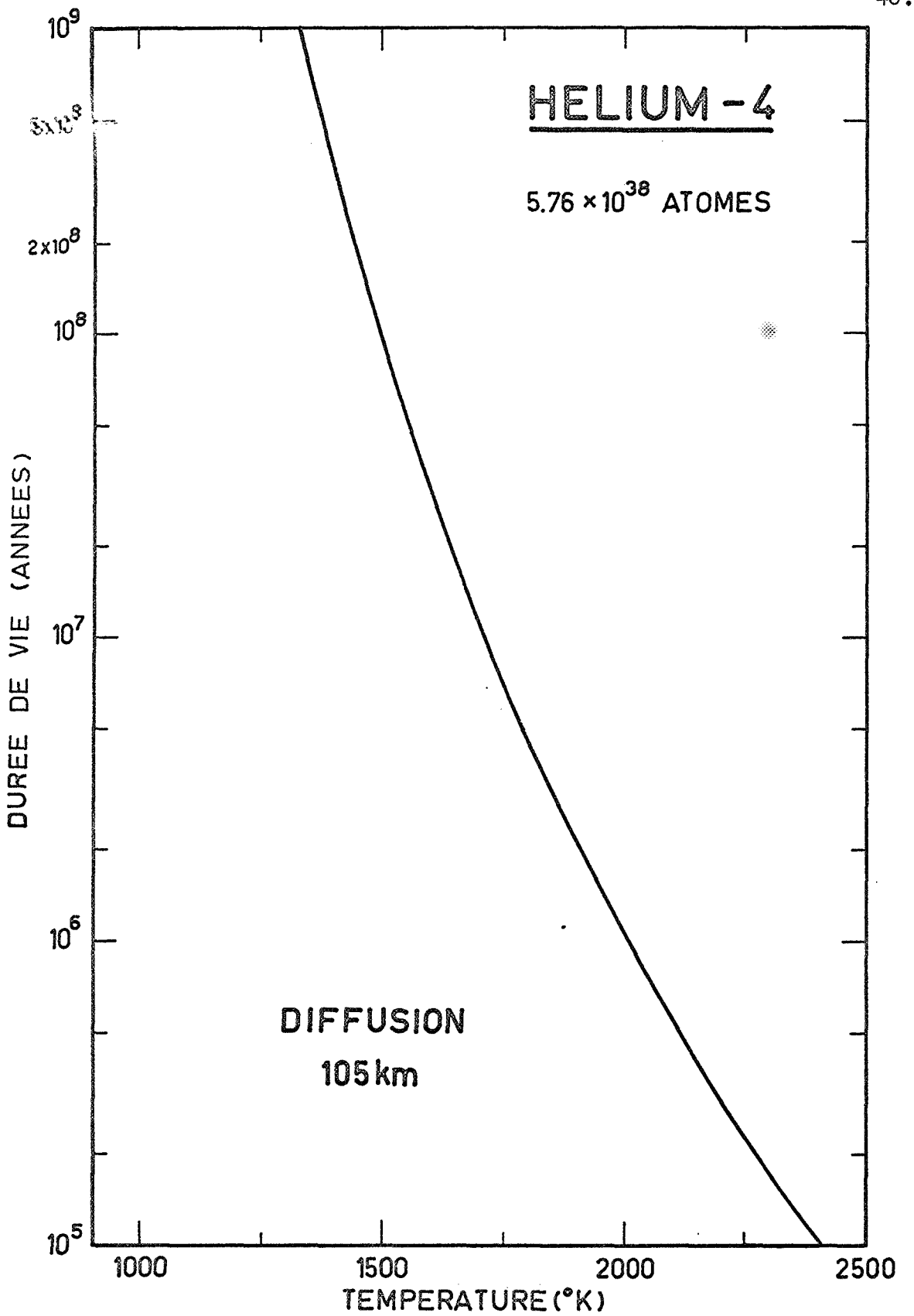

Fig. 6.- Lifetime of atmospheric helium as a function of the thermopause temperature when the helium-4 diffusion level is taken at 105 km.

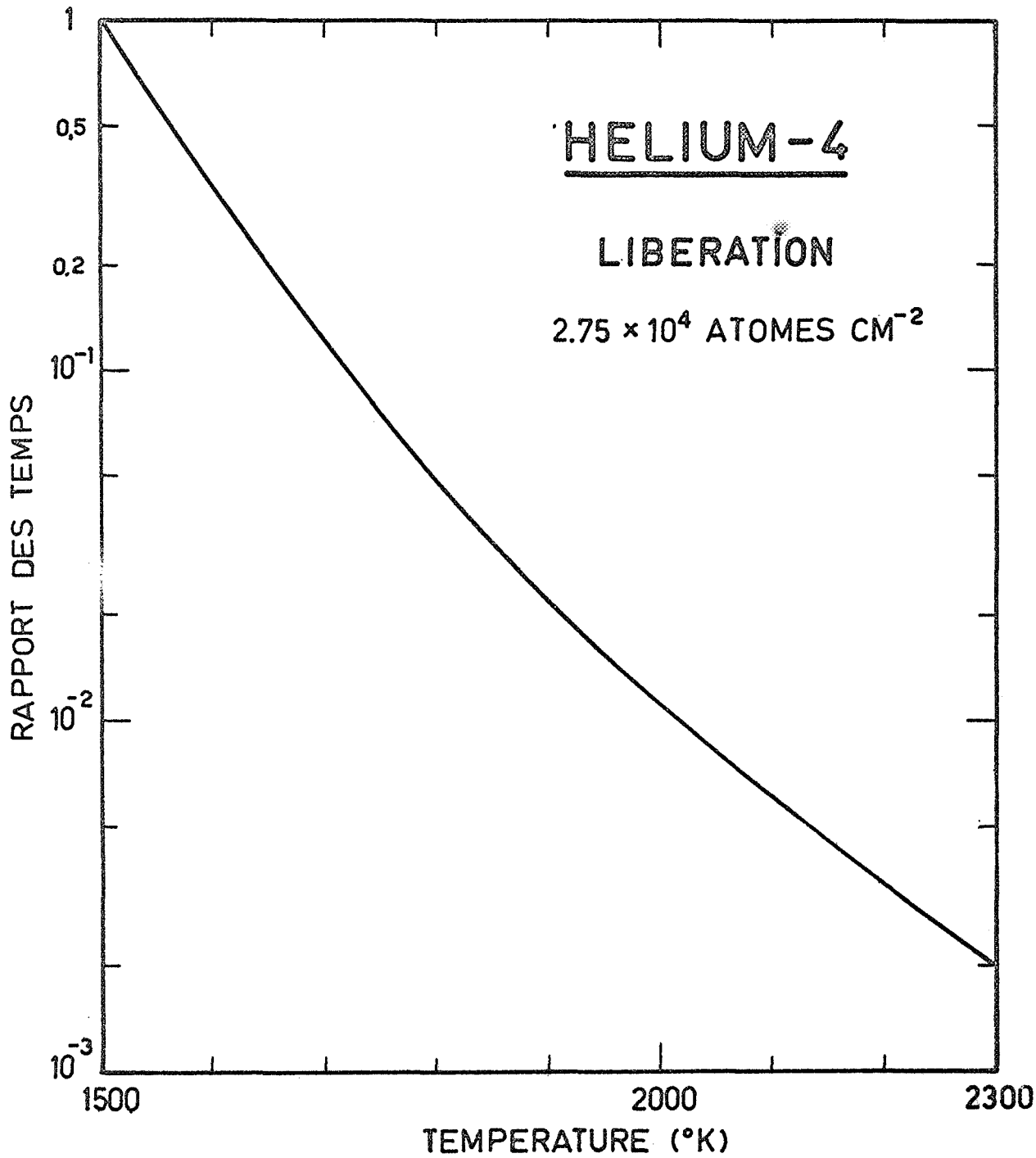

Fig. 7.- Effect of thermopause temperature on the escape of a constant number of atoms He^4.

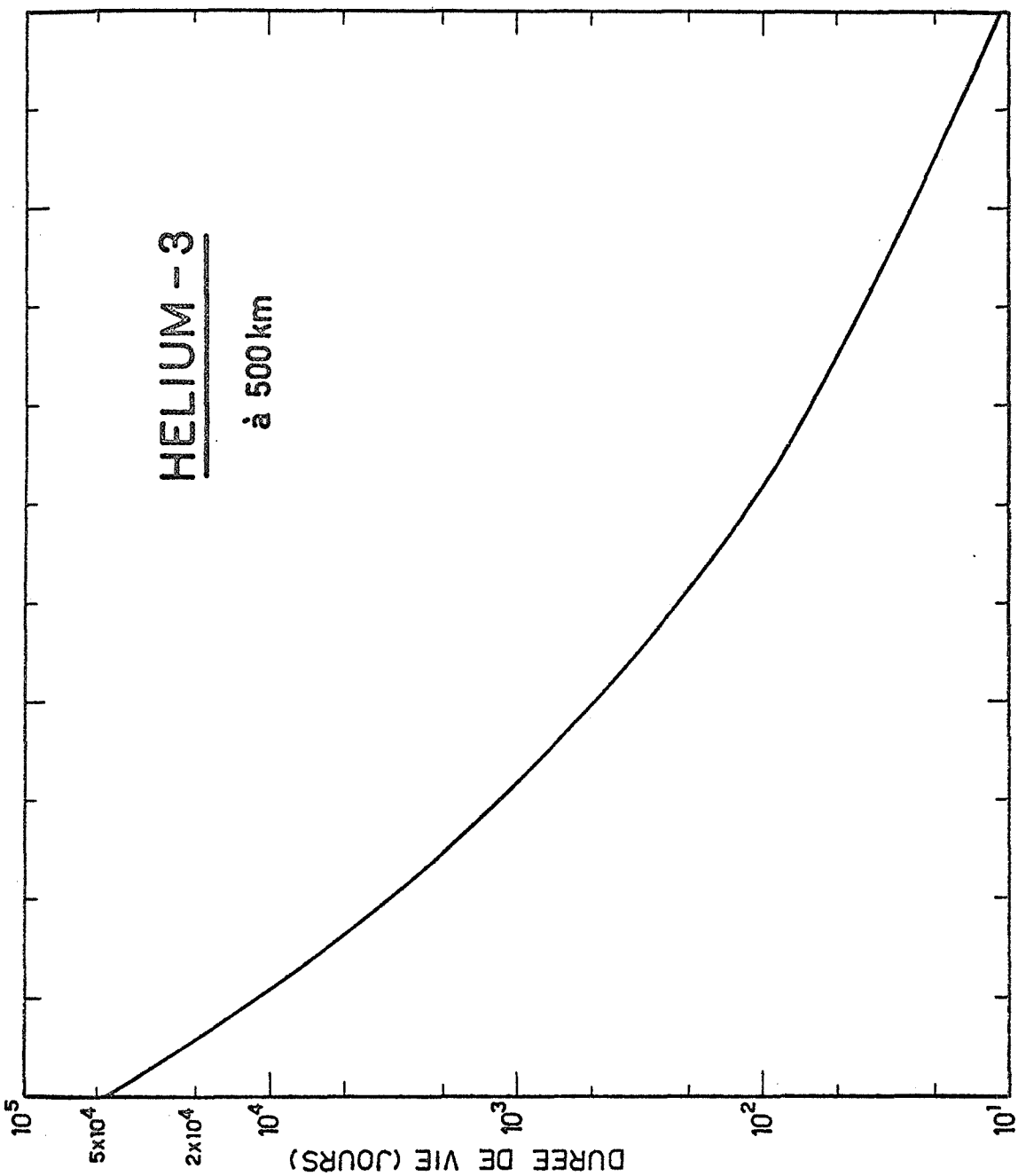

Fig. 8.- Durée de vie d'un atome d'hélium-3 à 500 km en fonction de la température de la thermopause. (Lifetime of a helium-3 atom as a function of thermopause temperature).

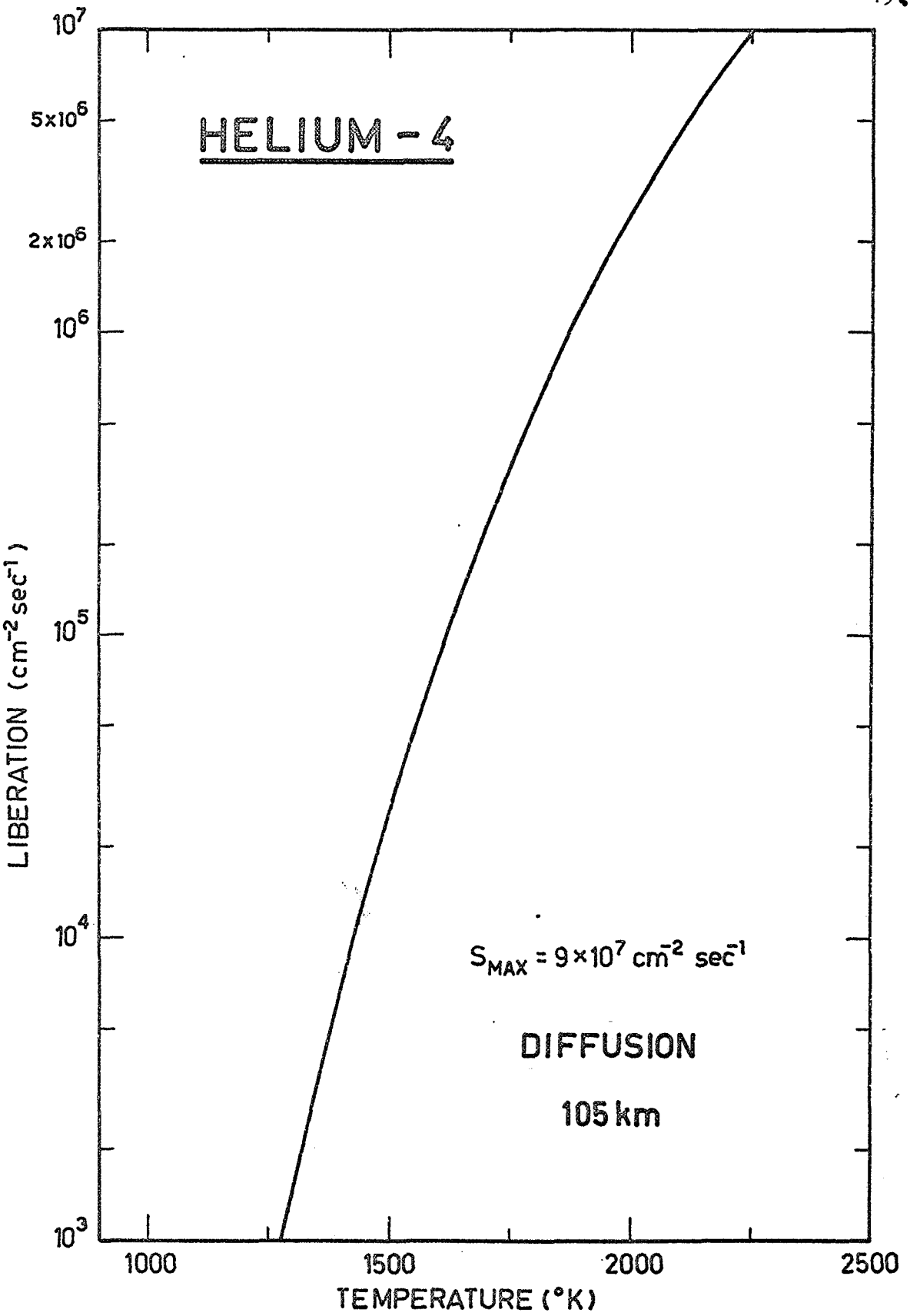

Fig. 9.- Flux of He^4 atoms liberated per cm^2 and per second as a function of thermopause temperature when the equilibrium level of diffusion is placed at 105 km.

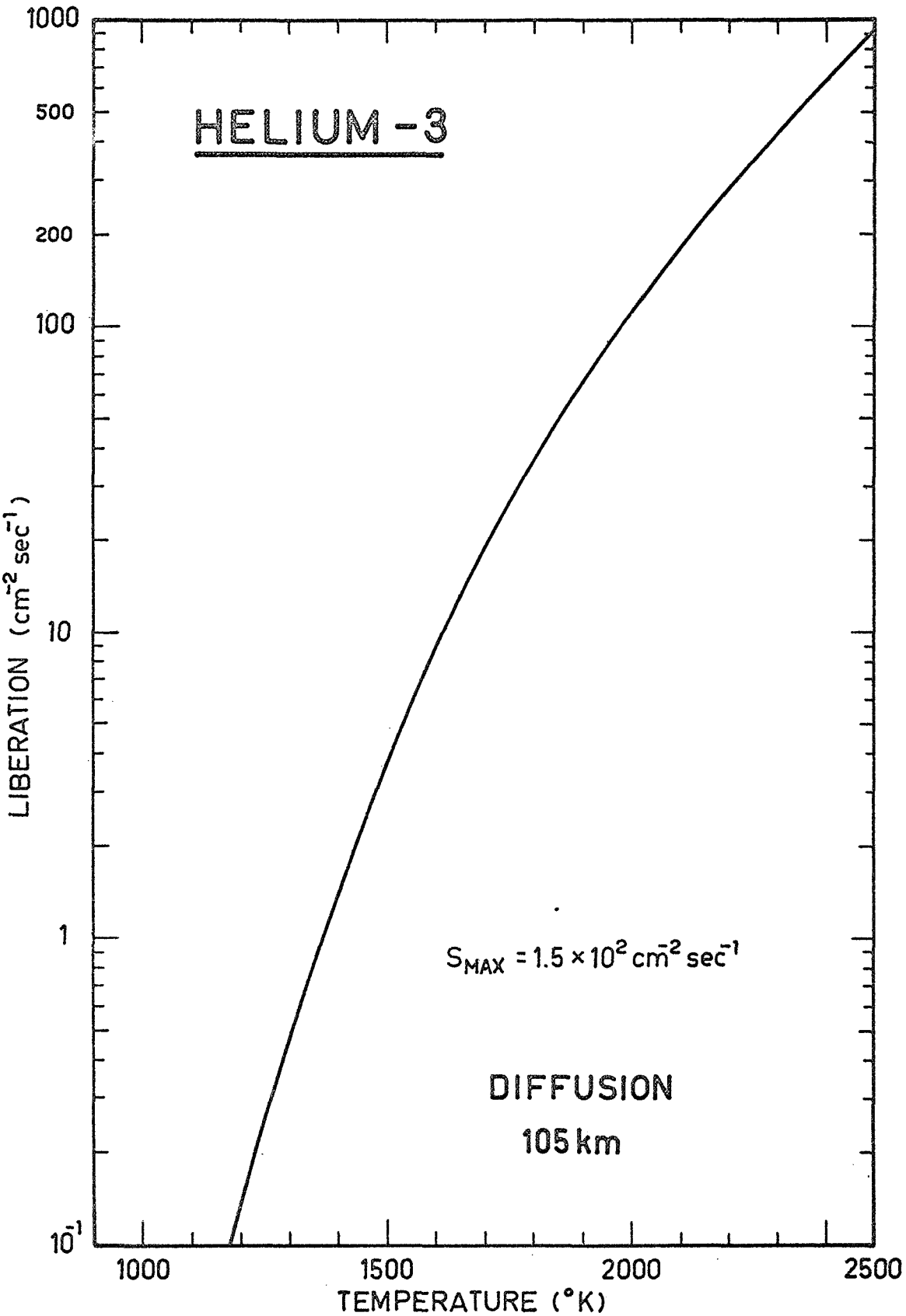

Fig. 10.- Flux of He^3 atoms liberated per cm^2 and per second as a function of thermopause temperature when the beginning of diffusion equilibrium is placed at 105 km.

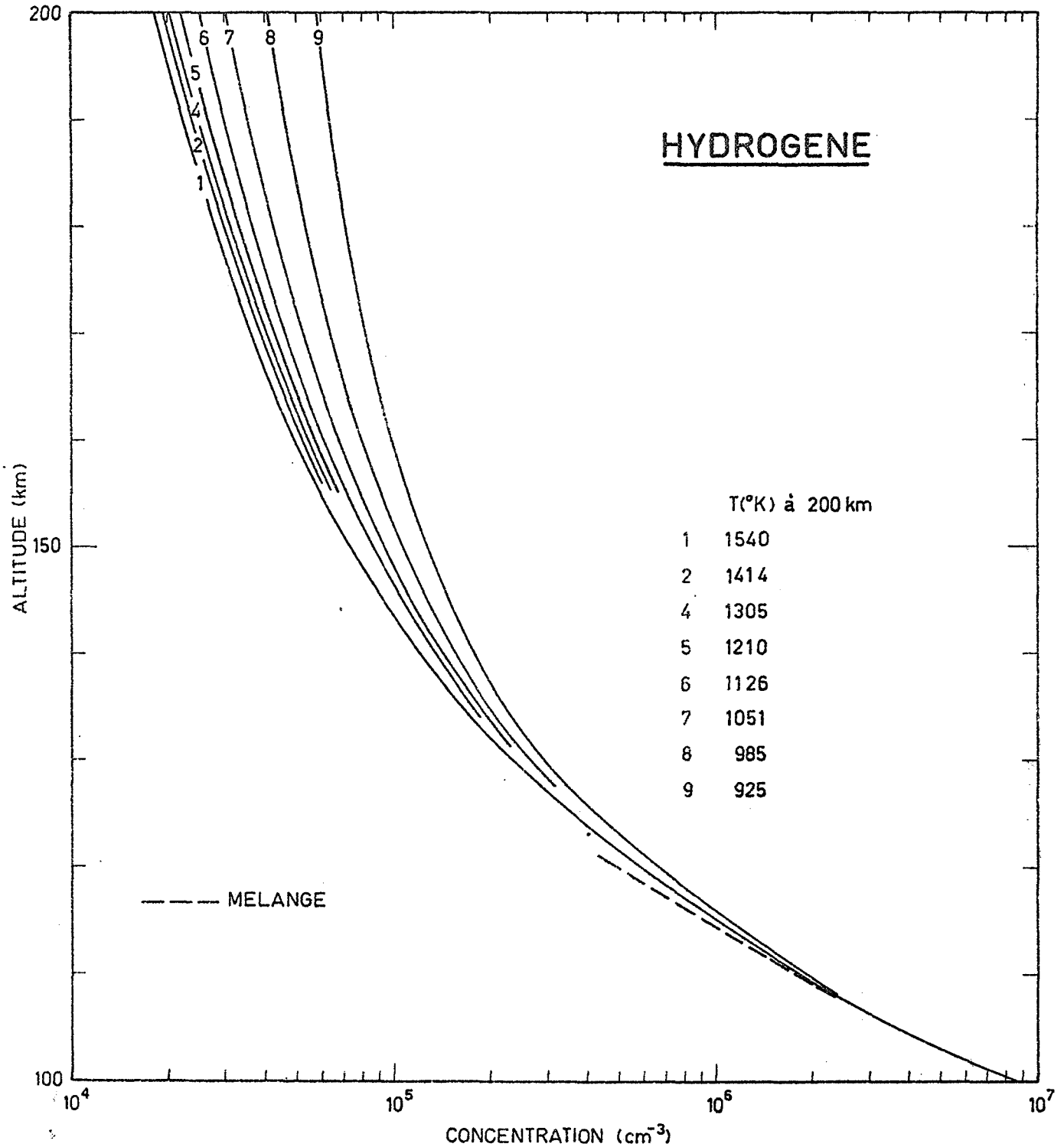

Fig. 11.- Vertical distribution of the concentration of atomic hydrogen when the diffusion flux at 100 km is 2.5×10^7 atoms $cm^{-2} sec^{-1}$. It may be noted that below 120 km, the distribution is not very different from that corresponding to mixing.

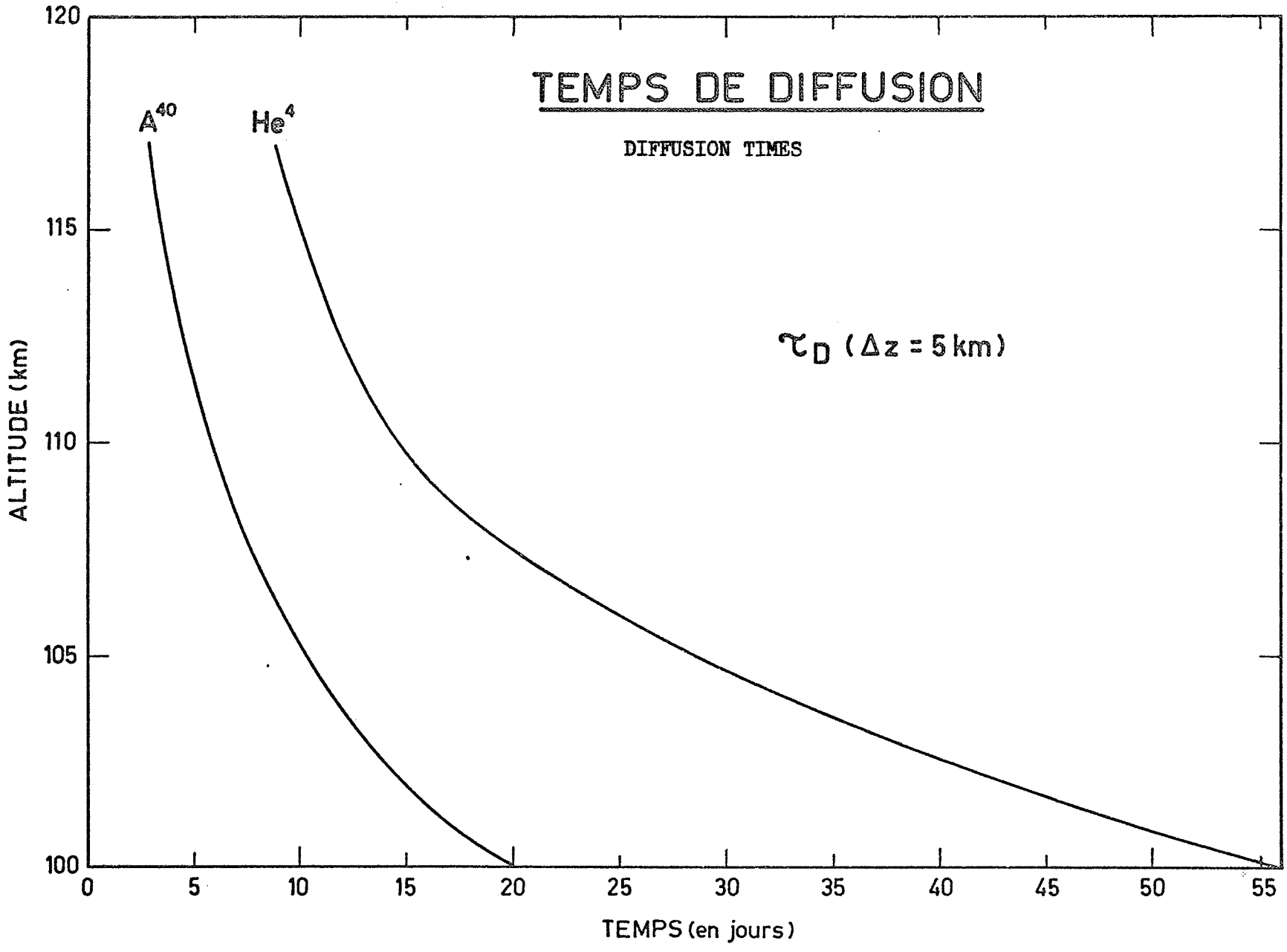

Fig. 12.- Temps nécessaires pour abaisser de 5 km le début du niveau de diffusion de l'argon et de l'hélium.
Times necessary to bring down by 5 km the beginning of the diffusion level for argon and helium.

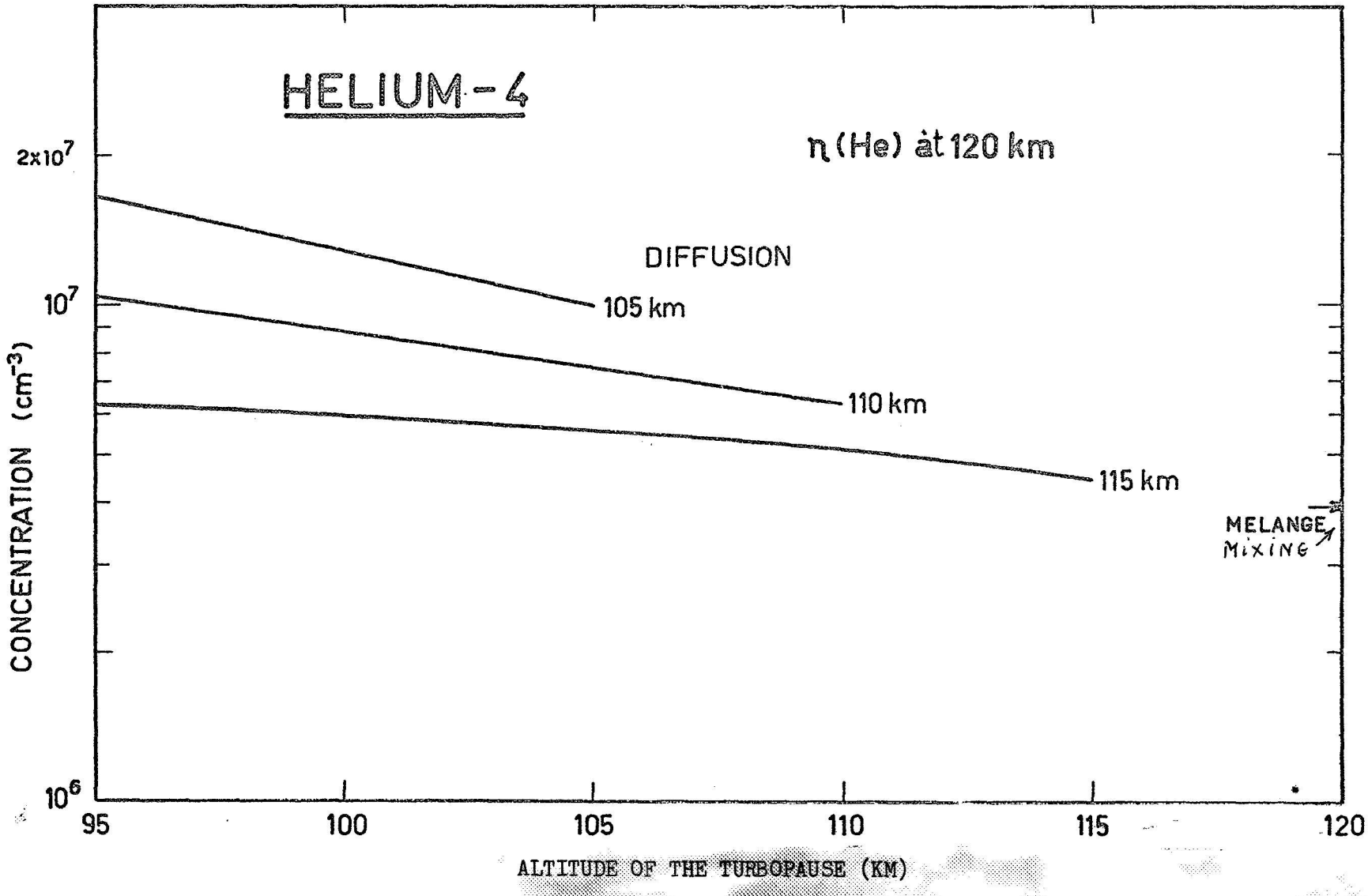

Fig. 13.- Concentrations of helium at 120 km taking simultaneously into account the level of the turbopause and that of the diffusion equilibrium

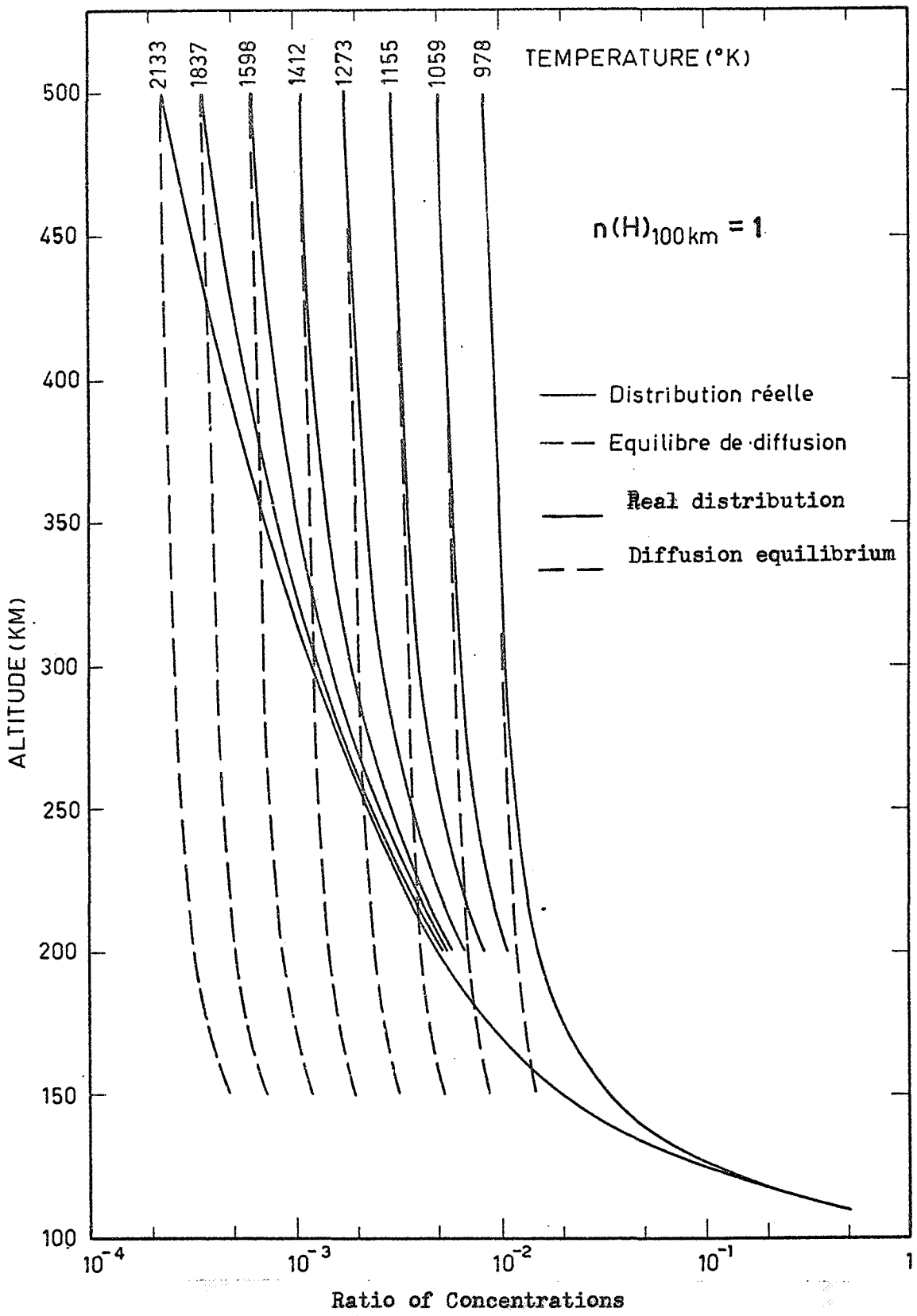

Fig. 14.- Comparison of vertical distributions of the hydrogen atom in case of a real distribution and in case of distribution resulting from the diffusion equilibrium.

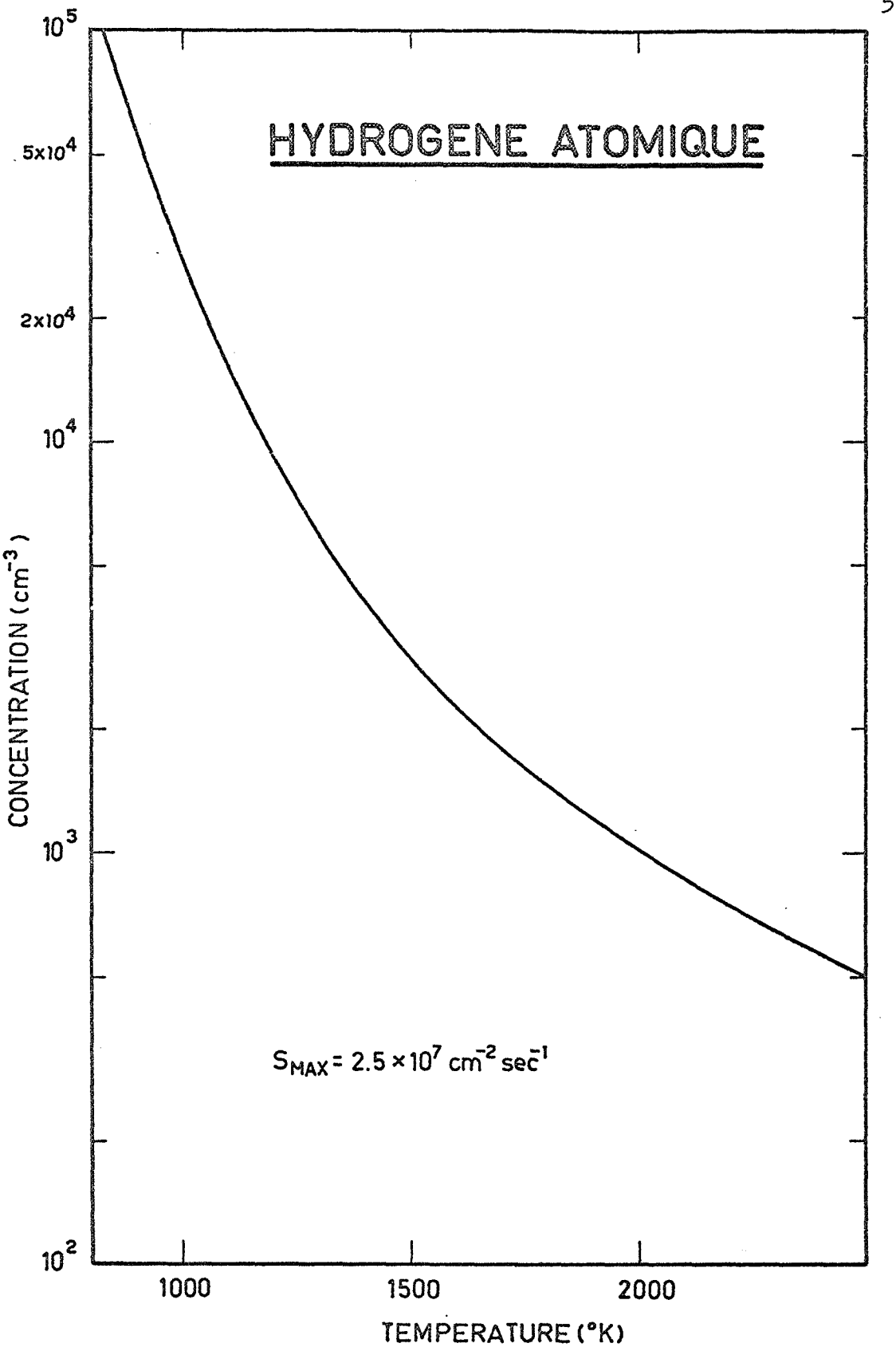

Fig. 15.- Concentration of atomic hydrogen at 500 km as a function of temperature of the thermopause when the diffusion flux at 100 km is $S = 2.5 \times 10^7$ atoms cm^{-2} sec^{-1}.

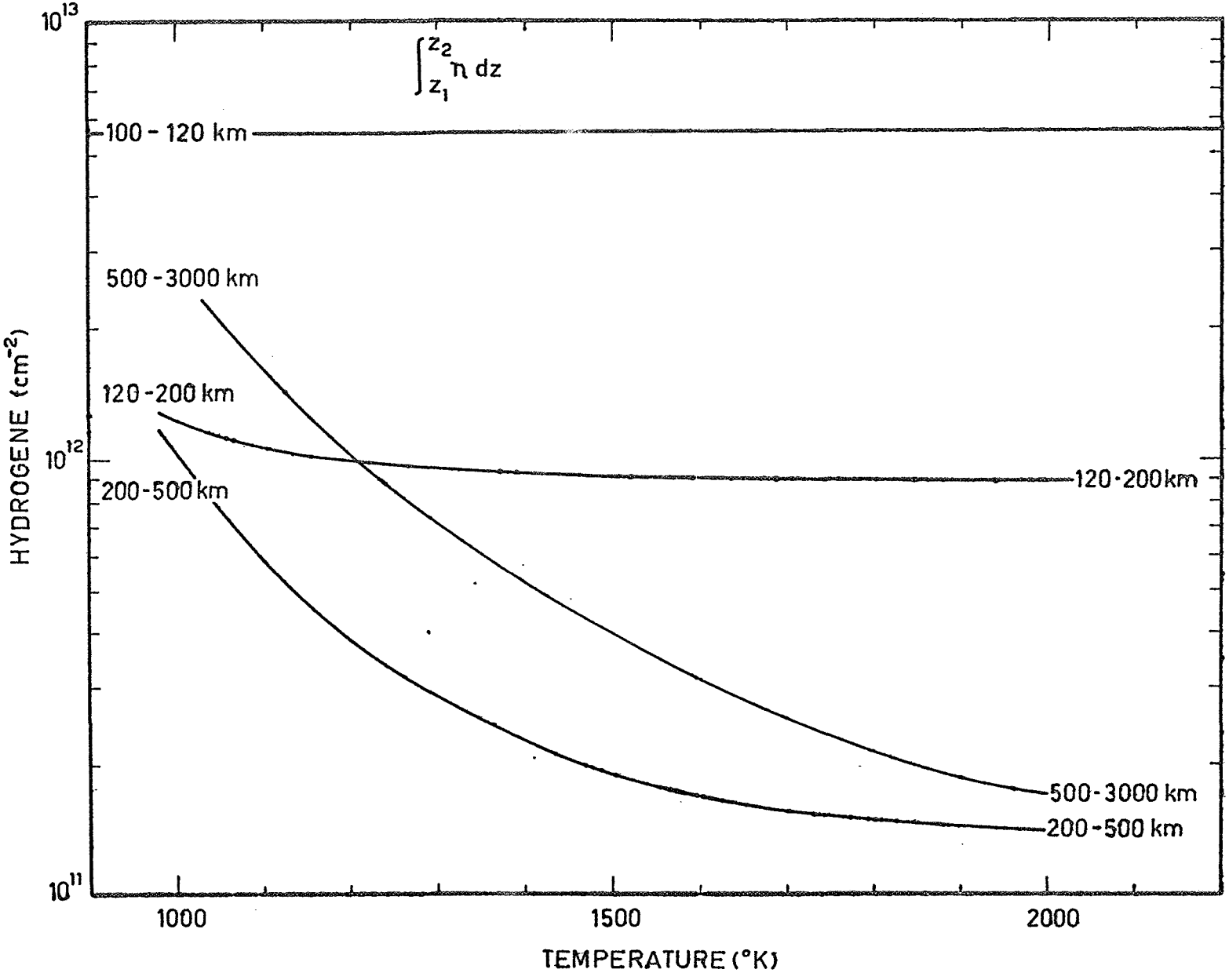

Fig. 16.- Nombre total d'atomes d'hydrogène dans des colonnes comprises entre 100 km et 120 km, 120 km et 200 km, 200 km et 500 km, 500 km et 3000 km. (Total numbei of hydrogen atoms in columns between 100 km and 120 km, 120 km and 200 km, 200 km and 500 km, 500 km and 3000 km).

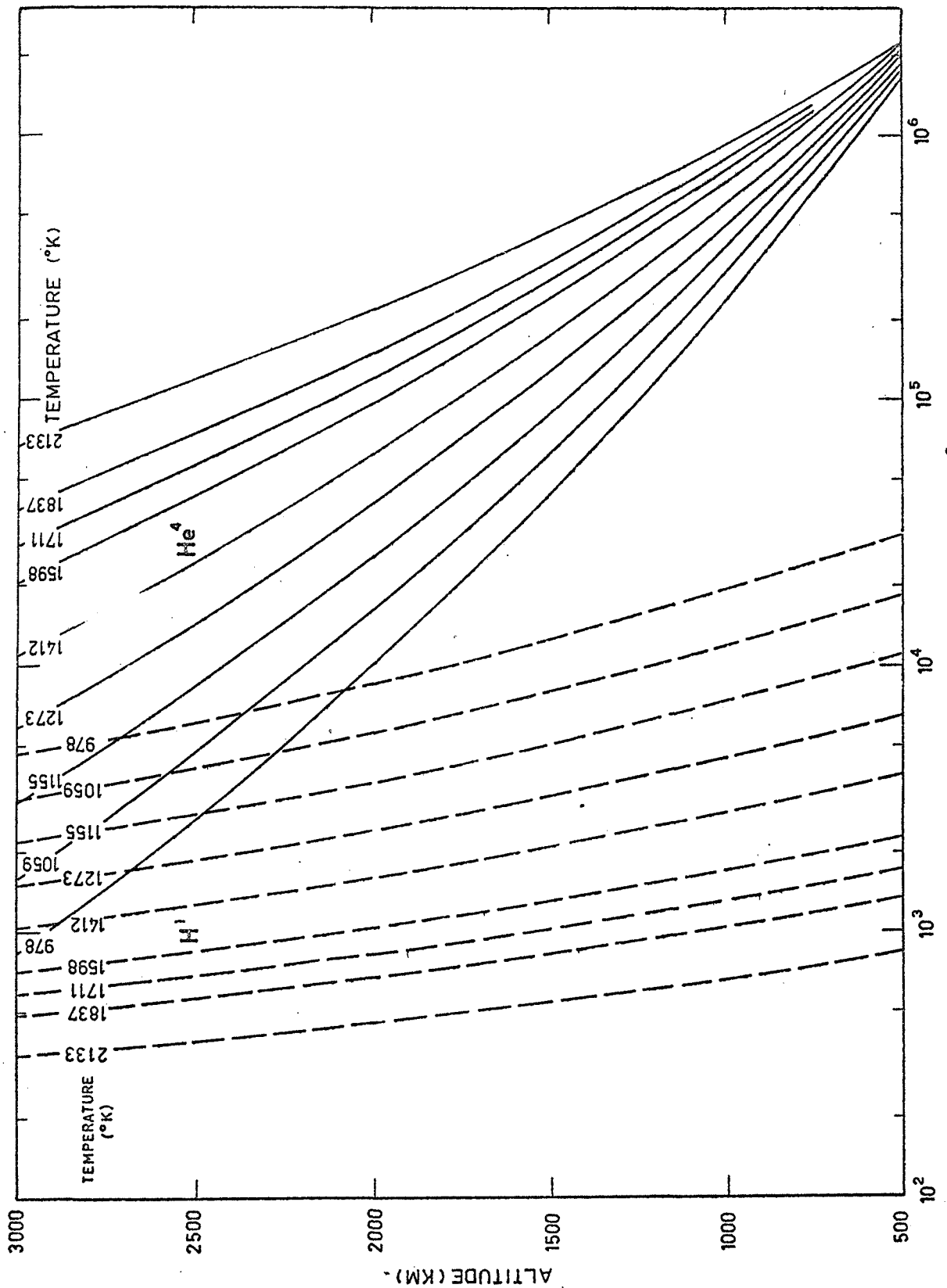

Fig. 17.- Distribution verticale des concentrations de l'hydrogène atomique et de l'hélium en fonction de la température de la thermopause. (Vertical distribution of atomic hydrogen and helium concentration as a function of thermopaus temperature.)

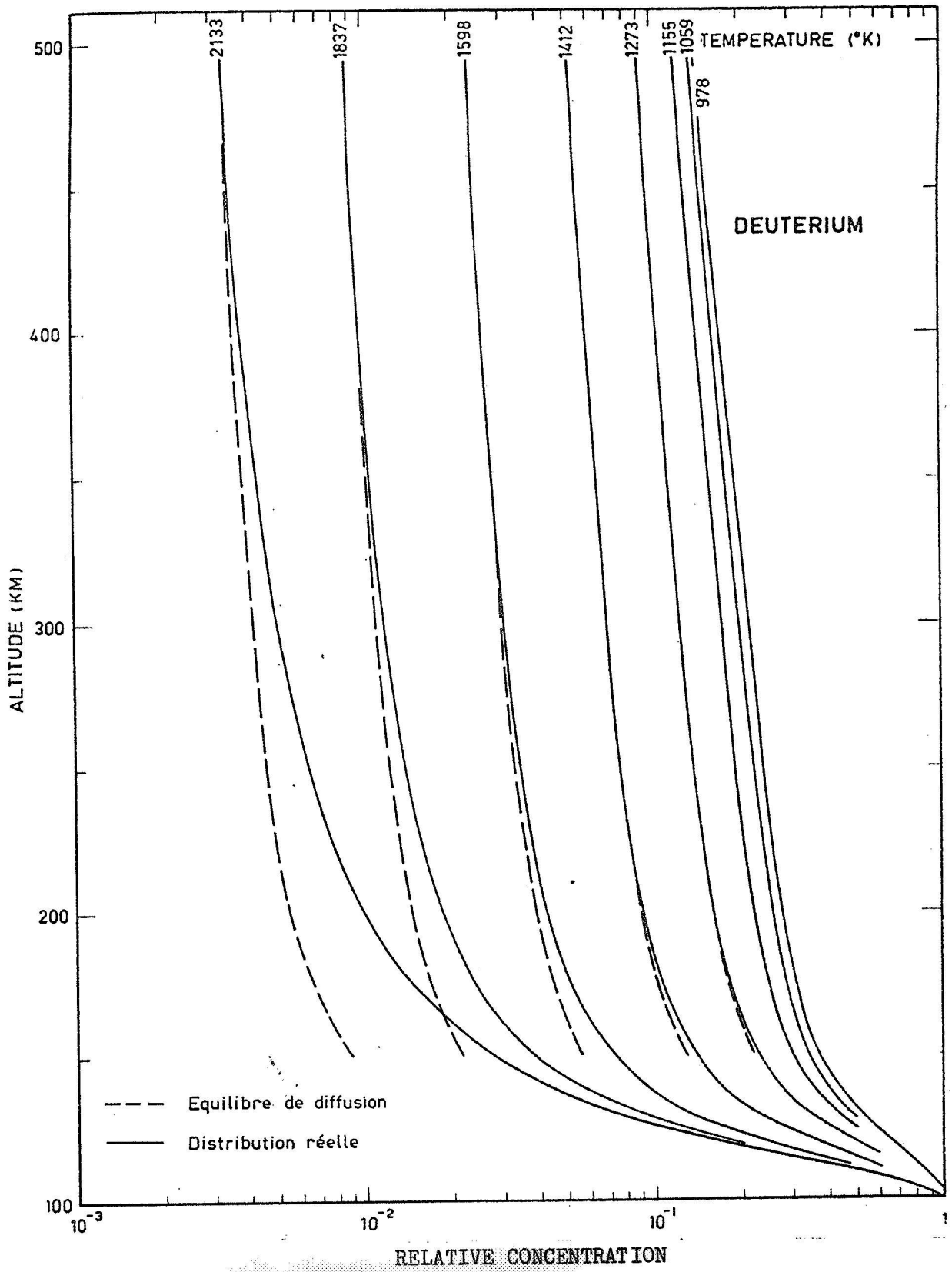

Fig. 18.- Relative deuterium concentration in true vertical distribution compared to diffusion equilibrium extrapolation beginning at 500 km

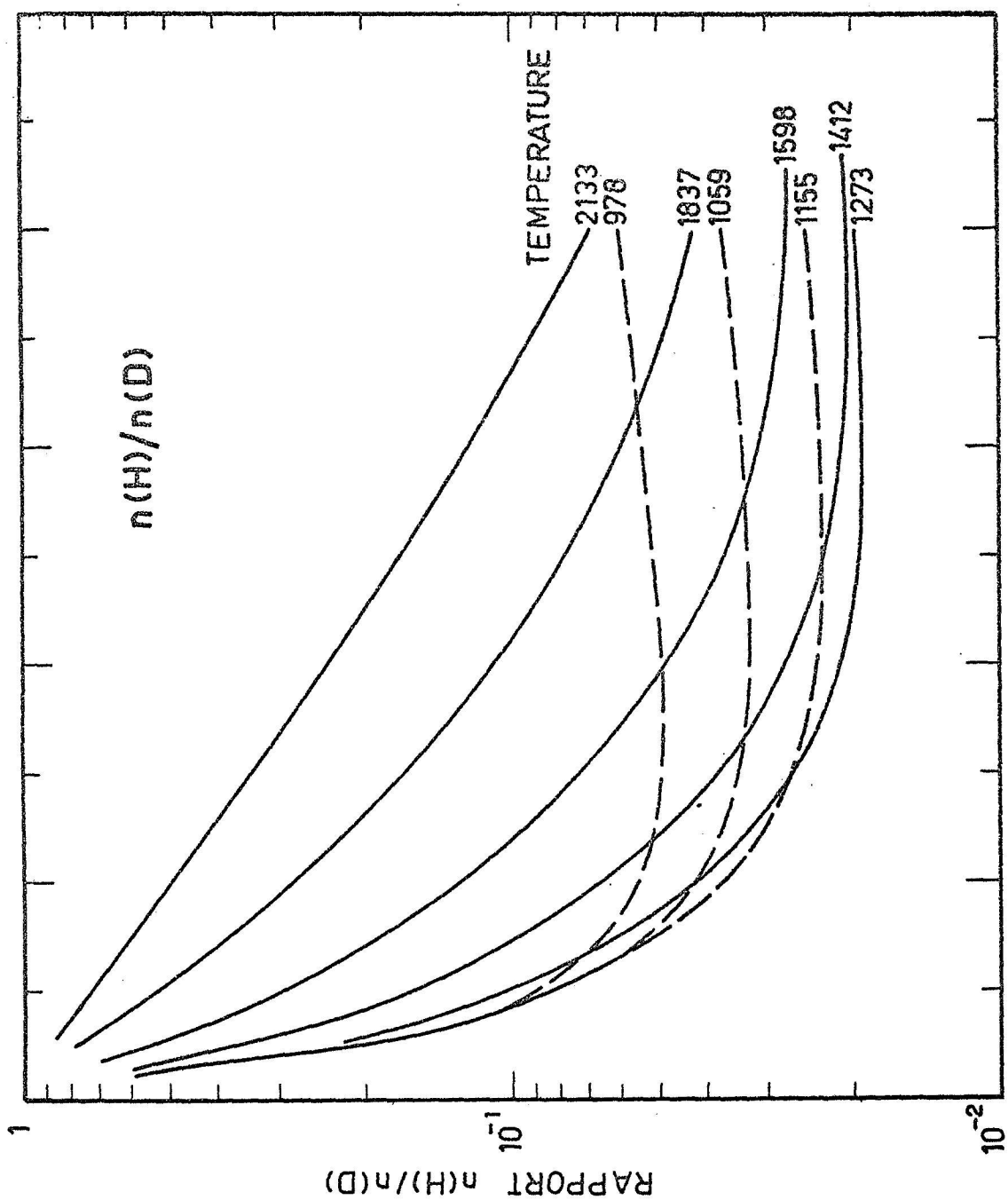

Fig. 19. Decrease in the ratio n (H) / n (D) from 100 km to 500 km. The value relative to 100 km being the unity and the indicated temperatures being those of the thermopause.

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